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TRACE ELEMENTS IN PYRITE, PYRRHOTITE AND CHALCOPYRITE OF DIFFERENT ORES

J. E. HAWLEY AND IAN NICHOL

ABSTRACT

Trace element determinations on pyrite, pyrrhotite and chalcopyrite from magmatic nickel ores, and hydrothermal copper and gold ores indicate that (1) certain elements, especially cobalt and silver, and to a lesser extent nickel and lead, show a preferential concentration in the three sulfides; (2) that some significant variations occur in deposits of different type and in different deposits of similar type, and (3) that there is a striking qualitative similarity in trace elements of all the deposits. Possible interpretations are given. These favor magmatic sources for the ores and suggest local contamination of rising ore fluids as a cause of some of the variations.

INTRODUCTION

JUST as trace elements have been studied in various igneous rocks to determine their distribution in various silicates that crystallize successively, so it is of interest to study sulfide-bearing ores to see if there is any systematic distribution of minor elements in the common sulfides such as pyrite, pyrrhotite and chalcopyrite, minerals common to several types of ore deposits, and minerals that appear time and again to have crystallized successively during a single period of mineralization. That such a study might be useful was indicated by Fleischer in 1955.

Samples of such sulfides from three types of deposits (1) magmatic nickel ores, (2) hydrothermal copper ores and (3) pyritic gold ores have been concentrated and analyzed for their minor constituents. Most of these samples were also analyzed for their selenium content, results of which have already been presented. The following paper deals with other minor constituents of such ores.

From the literature it has already been indicated that

(1) A variation in trace elements seems to occur between sulfide deposits of varying type-magmatic, hydrothermal, or sedimentary.

(2) A variation in the Co: Ni ratio in pyrite possibly reflects a difference in the partition of elements in varying types of deposits, which may be related to different temperatures of formation or host rocks.

(3) There appears to be a definite partition of Co and Ni between contemporaneous pyrite and pyrrhotite—Co is enriched in pyrite and Ni in pyrrhotite.

(4) A variation in the Co and Ni content of pyrite has been observed in several mines where a depth study has been carried out. This has also been noted in pyrrhotite in one area; but later workers have discredited this conclusion.

(5) Variables such as grain size, host rock or matrix, or distances from intrusives have all been considered but more work is needed on such topics before any generalizations can be made.

MODE OF OCCURRENCE AND SOURCE OF TRACE ELEMENTS IN SULFIDES

The trace elements in sulfides as noted by Auger may be present in a variety of ways—admixed solid inclusions, fluid inclusions, various types of substitutional, interstitial or "proxy" solid solutions, and some finely intergrown ex-solution products. Factors governing solid solution have been described adequately and will not be repeated here.

Many of the trace elements may have had their source in whole or in part in the ore fluid from which the sulfides have been deposited. Under equilibrium conditions if only one ore fluid was involved, one might expect uniform ratios of trace elements in any one mineral. But this is seldom likely to be the case. In many sulfide ores there is abundant textural evidence that ore fluids have varied in composition with time and the sulfides represent not simply crystals that have grown from a solution but replacements of a variety of wall rocks and also of earlier sulfides. The degree of contamination with solid inclusions in general may be expected to increase with the amount of replacement but more directly with those factors that inhibit complete replacement. Contamination may also occur by the incorporation in the replacing crystal lattices of those elements of favorable size factor, valency, and chalcophile character, which were formerly in silicates or oxides of wall rocks. It is clear, therefore, interpretations of trace element distribution in sulfides must be made with these points in mind.

Elements for which quantitative analyses have been made are: Ag, Co, Cr, Ni, Pb, Sn, Ti and V. In addition more sensitive qualitative tests were made and such elements as Au, As, Bi, Mn and Mo were noted. While many of these have ionic radii close enough to that of Fe^{+2} for some isomorphous substitution to occur in the sulfide lattice, Co and Ni are probably best accounted for in this way. Silver with a radius too large to easily substitute for Fe^{+2} Cu^{2+} or Cu^{+} none the less is clearly concentrated in the copper sulfide and undoubtedly proxies for Cu^{+}. Chromium, titanium and vanadium with higher valencies than Fe^{2} may to a limited extent proxy for Fe. Erratic

distribution of these even with one sample suggest they may be present in inclusions. Constant ratios as in samples from Sudbury, however, suggest a more intimate association of these with the lattices of the sulfides but also may simply reflect uniformity of inclusions present.

Lead, like silver has too large an ionic radius to substitute to any extent for Fe^{+2} and its rather erratic occurrence in the different sulfides strongly suggests the lead determinations are in large part due to small inclusions of galena, although, as will be shown, lead appears to favor pyrite over pyrrhotite and chalcopyrite in four of six copper ores.

Tin, on the basis of its ionic radius is acceptable in the sulfide lattice and where present in constant amounts may well be present in this form. In some cases, (Manitouwadge) tin was found in rather high amounts in pyrite (0.7%) and chalcopyrite (0.35%). Careful study of these samples showed the presence of admixed cassiterite (confirmed by X-ray analysis) and it is not possible to say in this case what proportion, if any, is present in the lattice.

Preparation of Samples.—Samples of the sulfides were first crushed in a Plattner mortar and then sieved in a set of Tyler screens. The fraction (− 100 + 150) mesh was retained and purified. Quartz and gangue were floated off using tetrabromoethane, and the sample was then washed with acetone and dried under a heat lamp. The several sulfides were then separated by means of an S. G. Frantz Isodynamic Separator. The purification procedure was carried out with constant resort to a binocular microscope to determine that the various fractions were pure.

Purity of Samples.—In cases where there was any doubt as to the purity of the pyrite or chalcopyrite sample, the sample in question was tested on an X-ray spectrograph where its Fe or Cu content was compared to that of a sample known to be very pure. In cases where the Fe or Cu fell more than 1 percent below that of the standard the sample was rejected but most of the samples tested were of very high purity.

Spectrographic Conditions.—The spectrographic conditions found most suitable with a 2m (24,400 lines to the inch) spectrograph, a Stallwood air-jet, and a d.c. arc discharge from an Applied Research Laboratories Multisource Unit were as follows:

	Various metals	Silver
Time	30 secs	20 secs
Gap	3 mm	3 mm
Shutter	25	25
Amps	8.5	9.0
Volts	300	340
Transmission	100%	4%
Grating doors	0.2–0.3	0.2–0.4
Filter	3	—
Electrodes	Sample: 1/4" center post electrodes with 1/8" crater undercut. Counter: 1/4" sharpened with pencil sharpener.	
Polarity	Sample positive.	
Film	Spectrum analysis Film No. 1	
Photography & Densitometry	Standard methods used throughout. A.R.L. Densitometer and calculating board for detmn. of Intensity ratios.	

Preparation of Standards.—Standards were prepared by adding a mixture of 13 oxides or 'spex mix' to bases of pyrite, pyrrhotite or chalcopyrite and then diluting with further additions of the base. By such a procedure sets of standards ranging from 0.0008–5 percent of the element were made. The bulk of the samples had concentrations of traces below 0.1 percent but a higher range was required for high concentrations of tin and nickel in certain suites of samples. Three sets of standards using pyrite, pyrrhotite and chalcopyrite as bases were prepared for the range 0.0008–0.1 percent.

Line Pairs and Ranges of Usage.—The Sr line 2931.83 was used as the internal standard in all cases except in samples with high amounts of silver which were run under special conditions using In 3256.09 as the internal standard.

Element lines as used over specified ranges are given below:

TABLE 1

LINE PAIRS AND RANGES OF USAGE

Line	Percent Metal
Ag 3280.68 (Sr)	0.00006–0.002
Ag 3280.68 (In)	0.002 –0.1
Co 3433.04	0.003 –0.15
Cr 4254.34	0.0004 –0.007
Ni 2992.60	0.4 –7.0
Ni 3050.82	0.007 –0.1
Ni 3414.77	0.0003 –0.007
Pb 2833.07	0.007 –0.1
Sn 2429.50	0.3 –1.0
Sn 2839.99	0.001 –0.3
Ti 3371.45	0.004 –0.08
V 3185.40	0.002 –0.08

Working Curves.—The curves as established for Co and Sn using the Stallwood air-jet are shown in Figure 1. It is especially noted that standards prepared using either bases of pyrite, pyrrhotite or chalcopyrite give the same intensity ratio value for like standard concentrations in all cases except silver. This is a striking example of the value of the air jet in that a variation in concentration of the major constituents has no effect on the sensitivity of the minor elements in such minerals.

Precision or Reproducibility.—Reproducibility as distinct from accuracy is a measure of the ability of a method to reproduce a determination. The precision of the study was checked throughout the course of the work by running sets of check standards every two or three weeks over a period of six months and thus checking the working curves. In all cases the precision was very good and the greatest difference was 4 to 5 percent. Cobalt determinations using different cobalt lines (3433 and 3395) differed on the average by not more than 5.5 percent of the amount present.

Accuracy.—To check the accuracy of the spectrographic work, as no well analyzed material was available, a series of samples were run on a Phillips X-ray spectrograph (Tungsten tube) for cobalt. The results obtained are compared with those from the spectrographic analysis. The standards for

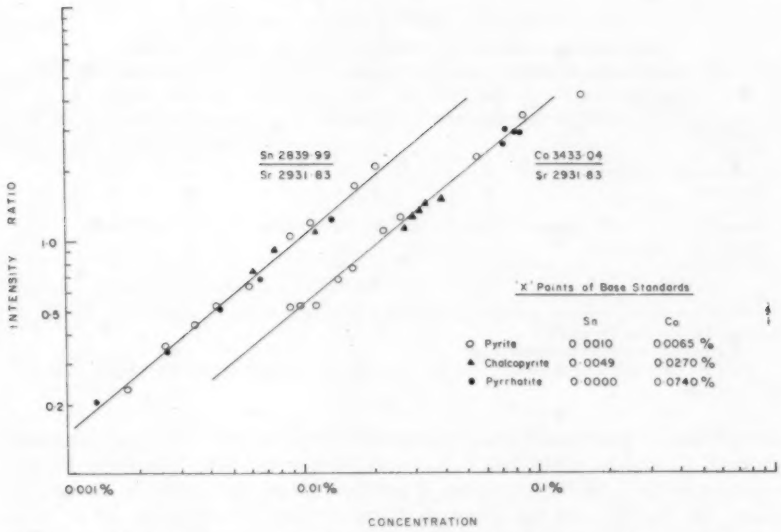

FIG. 1. Working curves for tin and cobalt using the three sulfides as base standards as determined with Stallwood air-jet.

the X-ray spectrograph were the same as those used in the spectrographic analyses.

TRACE ELEMENT CONTENT OF INDIVIDUAL MINERALS

Pyrite, pyrrhotite and chalcopyrite are dealt with in turn with respect to varying trace element contents in three different types of ores. Variations with depth and type of pyrite are also considered. It is evident in the detailed discussion that marked variations do occur.

PYRITE

Qualitative Analyses.—In addition to the elements studied quantitatively in pyrite from the various deposits, (i.e. Co, Ni, Ag, Cr, Pb, Sn, Ti, V) qualitative analyses were made for others such as Au, As, Bi, Mn, Mo and also for Sn and V where amounts present were not determinable by the quantitative procedure. Sudbury pyrites are known to be of at least two ages, one earlier

TABLE 2

COBALT DETERMINATIONS ON SPECTROGRAPH AND 'X-RAY'

Sample No.	% Co (spec.)	% Co X-ray
N48	0.100	0.090
N55	0.110	0.110
N64	0.110	0.128
N67	0.150	0.135
N125	0.052	0.050

TABLE 3

QUALITATIVE ANALYSIS OF PYRITES FROM THE VARIOUS MINES

Location	Au	As	Bi	Mn	Mo	Sn	V
Sudbury*				x			x
Flin Flon	x	x		x	x	x	x
Chibougamau (Campbell)	x	x	x	x	x	x	x
Quemont	x		x	x	x	x	x
Noranda (+)	x		x	x	x	x	x
Normetal	(x)	x	x	x	x	x	x
Manitouwadge (Geco)		(x)	x	x	x	x	x
McIntyre	x	x	x	x	x	x	x

* Eight Sudbury pyrites, according to Lewis (1950) were found to contain Co, Ni, Zn, Al, Cu, Cr, Mg, Pt, Ag, Pb and V.

(x) Present in only one sample.

(+) Pd, .012 g/t and traces Pt and Rh detected in Noranda pyrite (Hawley & Rimsaite, 1953).

than pyrrhotite-pentlandite-chalcopyrite and the other later. Early pyrite shows traces of Pt and is richer in Co, whereas later pyrite does not show Pt and is poorer in Co. Sudbury pyrites of pre-pyrrhotite age appear to be lacking in Au, As, Bi, Mo and Sn, although discrete minerals containing all of these except Mo are known to occur in the ores. Co, Ni, Cr, Ti, Ag, Mn, and Pb are present in all.

In the copper and gold ores, pyrite contains Co, Ni, Cr, Ti, V, Mn, Mo, Sn, Ag and, with some exceptions Au, As and Bi. Bi is missing or was not detected in only one copper ore (Flin Flon), As was not detected in the Noranda and Quemont pyrite, and Au likewise was not detected in Manitouwadge pyrite nor in all but one sample from Normetal.

TABLE 4

Mean Concentration of Trace Elements in Pyrites from Various Mines

Location	No. Specs.	Co%	Ni%	Co:Ni	Ag%	Cr%	Pb%	Sn%	Ti%	V%
Nickel Ores - (Misc. Sudbury)	8 5 1	1.33 1.03 0.18	.25 .10 .20	5.3 10.3 0.9	 .0024 .0002	.016	.017 .012	n.d.	.028	.0055
Copper Ores-										
Flin Flon	3	0.066	.0057	11.6	.0066	n.d.	.128	n.d.	.017	n.d.
Chibougamau (Campbell)	4	0.30	.011	27.3	.0039	.0012	.0020	n.d.	.030	n.d.
Quemont	4	0.084	.0022	38.2	.0055	.0082	.046	.021	.015	n.d.
Noranda	8	0.116	.0041	28.3	.0031	.0016	.020	.0058	.0096	n.d.
Normetal	12	0.062	.0016	38.7	.0042	.0012	.064	.022	.022	n.d.
Manitouwadge (Geco)	8	0.031	.0011	28.2	.0028	.0012	.067	.29	.010	n.d.
Gold Ores										
McIntyre	21	0.079	.041	1.9	.00044	.002	.042	n.d.	.115	n.d.
(low)	8	0.035	.034	1.0	.00057				.145	
(high)	13	0.105	.044	2.4	.0004				.099	

QUANTITATIVE ANALYSES OF PYRITES

The mean concentrations of the several trace elements are shown in Table 4. The ranges of percentages of Co, Ni, Ag, Pb and Sn found in pyrite as well as in pyrrhotite and chalcopyrite of the several deposits are shown in Tables A, B, C, D and E.

TABLE A

Occurrence, Range, Mean Concentration and Partition Distribution of Cobalt in Sulphides

Location	Pyrite			Pyrrhotite			Chalcopyrite			Dist. of Co
	No.	Range %	Mean %	No.	Range %	Mean %	No.	Range %	Mean %	Py:Po:Cpy
Hardy Mine				10	0.0135-0.050	0.034				
Misc. Sudbury	5	0.92-1.81	1.03				28	0.003-0.086	0.00154	5.7:1:0.41
Fecunis Lake				14	0.032-0.057	0.039				
Flin Flon	3	0.002-0.185	0.066	1	0.15	0.15	4	0.066-0.082	0.028	0.44:1:0.086
Chibougamau	4	0:031-0.86	0.30	20	0.051-0.19	0.11	9	0.0084-0.145	0.049	4.2:1:0.38
Quemont	4	0.014-0.265	0.084	2	0.022-0.030	0.026	4	0.0045-0.0066	0.0055	3.2:1:0.21
Noranda	8	0.033-0.285	0.116	9	0.011-0.076	0.038	8	0.0044-0.020	0.0077	3.1:1:0.20
Normetal	12	0.006-0.17	0.062	8	0.0046-0.078	0.019	8	0.005-0.040	0.016	3.3:1:0.84
Manitouwadge	8	0.012-0.049	0.031	9	0.0048-0.0088	0.0072	11	0.0046-0.022	0.0088	4.3:1:1.2
McIntyre (Av)	21	0.0066-0.34	0.079	1	0.055	0.055	1	0.036	0.036	1.4:1:0.65
(Lo)	8	0.0066-0.115	0.035							
(Hi)	13	0.0081-0.34	0.105							

TABLE B

Occurrence, Range, Mean Concentration and Partition Distribution of Nickel in Sulphides

Location	Pyrite			Pyrrhotite			Chalcopyrite			Dist. of Ni
	No.	Range %	Mean %	No.	Range %	Mean %	No.	Range %	Mean %	Py:Po:Cpy
Hardy Mine				10	1.13-1.67	1.33				
Misc. Sudbury	5	0.04-0.90	0.10				28	0.0086-1.03	0.037	0.035:1:0.29
Fecunis Lake				14	1.1-1.6	1.34				
Flin Flon	3	0.0010-0.012	0.0057	1	0.019	0.019	4	0.00071-0.0018	0.0013	0.30:1:0.07
Chibougamau	4	0.0029-0.019	0.011	20	0.012-0.082	0.032	9	0.0036-0.024	0.012	0.31:1:0.38
Quemont	4	0.0012-0.0039	0.0022	2	0.0011-0.0057	0.0034	4	0.00057-0.0028	0.00065	0.65:1:0.19
Noranda	8	0.001-0.038	0.0041	9	0.0009-0.01	0.0016	8	0.00053-0.010	0.0014	2.6:1:0.88
Normetal	10	0.00068-0.0084	0.0016	8	0.00064-0.0015	0.00099	5	0.00074-0.00145	0.00098	1.60:1:1
Manitouwadge	8	0.00058-0.0027	0.0011	8	0.00045-0.0053	0.0027	7	0.00056-0.0011	0.00072	0.41:1:0.27
McIntyre (Av)	21	0.0048-0.12	0.041	1	0.012	0.012	1	0.0048	0.0048	3.4:1:0.4
(Lo)	8	0.0053-0.08	0.034							
(Hi)	13	0.0048-0.12	0.044							

Cobalt.—Cobalt in pyrites ranges from 1.33 to .031 percent. Sudbury pyrite of pre-pyrrhotite age † is richest, but later age pyrite is comparable with some of the hydrothermal copper ores. Of the copper ores, pyrite from the Campbell-Chibougamau deposit shows the most cobalt and that from the Geco mine the least. Pyrite from the McIntyre gold mine containing cobalt in amounts similar to some of the copper deposits shows three times the amount of cobalt in pyrite that gave high temperature effects at 450° to 600° C compared with pyrite showing low temperature effects at 125° to 200° C. Although doubt has been cast on the interpretation that these represent temperatures of formation, it seems clear there are two types of pyrite as far as cobalt content is concerned. In addition it may be noted cobalt

TABLE C

Occurrence Range, Mean Concentration and Partition Distribution of Silver in the Sulphides

Location	Pyrite			Pyrrhotite			Chalcopyrite			Dist. of Ag
	No.	Range %	Mean %	No.	Range %	Mean %	No.	Range %	Mean %	Py:Po:Cpy
Hardy Mine				10	0.00009-0.00023	0.00014				
Misc. Sudbury	5	0.0002-0.0003	0.00024				28	0.0091-0.034	0.018	1.6:1:120
Fecunis Lake				14	0.000068-0.00073	0.00016				
Flin Flon	3	0.0016-0.011	0.0060	1	0.008	0.008	4	0.0048-0.030	0.016	0.75:1:2.0
Chibougamau	4	0.00017-0.0082	0.0039	20	0.00015-0.0052	0.00091	9	0.0058-0.0172	0.0119	4.3:1:13.1
Quemont	4	0.0010-0.014	0.0055	3	0.0010-0.0022	0.0016	3	0.0102-0.0255	0.016	3.4:1:10.6
Noranda	7	0.00037-0.0098	0.0031	8	0.00011-0.0015	0.00085	8	0.0056-0.0122	0.0091	3.6:1:10.7
Normetal	11	0.00027-0.0185	0.0042	7	0.0008-0.021	0.011	8	0.021-0.069	0.031	0.38:1:2.8
Manitouwadge	8	0.000031-0.012	0.0028	7	0.000105-0.0175	0.0072	11	0.0105-0.044	0.030	0.39:1:4.2
McIntyre (Av)	19	0.000082-0.0911	0.00044	1	0.00009	0.00009	1	0.0165	0.0165	4.9:1:83.0
(Lo)	6	0.00017-0.0811	0.00057							
(Hi)	13	0.000082-0.00084	0.00040							

content in McIntyre samples of both "low" and "high" types show an increase in this element with depth. In contrast Noranda pyrite samples indicate a slight decrease at depth in cobalt as noted by Auger—whatever the reason.

Nickel.—Nickel content of pyrites range from 0.25 to 0.0011 percent. Those from Sudbury ores that are the highest in nickel as might be expected do not include nickelian pyrite which has been reported to contain from 4 to 6.5 percent nickel. Rather surprisingly pyrites of the copper deposits are in every case lower in nickel than that from the gold deposit. Of the two types of pyrite at McIntyre that showing higher cobalt is richer in nickel and in both types a slight increase with depth (to 6500′) is apparent as is the case

† Recent study by A. J. Naldrett has shown this early pyrite to be of two types, one corroded by pyrrhotite, the other in crystals in pyrrhotite, a likely reaction product with sulphur vapor. The former contains only 0.60% Co, but the latter is richer with 1.35% Co.

TABLE D

Occurrence, Range and Mean Concentration of Lead in the Sulphides

Location	Pyrite			Pyrrhotite			Chalcopyrite		
	No.	Range %	Mean %	No.	Range %	Mean %	No.	Range %	Mean %
Hardy Mine				0/10					
Misc. Sudbury	5/5	0.011-0.029	0.017				21/28	0.006-0.10	0.019
Fecunis Lake				14/14	0.0094-0.1	0.019			
Flin Flon	3/3	0.102-0.165	0.128	1/1	0.070	0.070	4/4	0.018-0.050	0.037
Chibougamau	2/4	0.0018-0.0021	0.0020	3/20	0.0022-0.074	0.033	6/9	0.0064-0.098	0.010
Quemont	3/4	0.0026-0.112	0.046	1/3	0.031	0.031	2/4	0.0066-0.0170	0.012
Noranda	4/8	0.013-0.035	0.020	3/9	0.011-0.037	0.020	3/8	0.0078-0.0165	0.0112
Normetal	12/12	0.012-0.115	0.0.84	8/8	0.0112-0.1	0.032	8/8	0.0074-0.051	0.023
Manitouwadge	8/8	0.032 0.12	0.067	7/9	0.015-0.1	0.037	9/11	0.007-0.052	0.026
McIntyre	6/12	0.025-0.058	0.042	0/1	n.d.		1/1	0.018	0.018

TABLE E

Occurrence, Range, and Mean Concentration of Tin in Sulphides

Location	Pyrite			Pyrrhotite			Chalcopyrite		
	No.	Range %	Mean %	No.	Range %	Mean %	No.	Range %	Mean %
Hardy Mine				0/10					
Misc. Sudbury	0/5						0/28		
Fecunis Lake				0/14					
Flin Flon	0/3			0/1			0/4		
Chibougamau	0/4			1/20	0.0023	0.0023	2/9	0.0013-0.0019	0.0016
Quemsont	2/4	0.019-0.023	0.021	0/3			4/4	0.0011-0.0062	0.0033
Noranda	5/8	0.0024-0.010	0.0058	5/9	0.0021 0.010	0.0071	8/8	0.0012-0.011	0.0043
Normetal	5/12	0.00115-0.084	0.022	4/8	0.00127 0.46	0.122	7/8	0.0019-0.011	0.0025
Manitouwadge	7/8	0.006-0.7	0.29	5/9	0.013-0.23	0.10	10/11	0.0021-0.35	0.10
McIntyre	0/21			0/1			0/1		

with cobalt. At Noranda nickel behaves as does cobalt, decreasing from .004 at or near surface to .0009 percent on the 31st level.

Cobalt-Nickel Ratios.—The cobalt-nickel ratios for the three types of deposits for the most part show distinct differences. Sudbury, 10; copper deposits, 11–38; and gold, 1–2.

With the exception of Flin Flon for which we had few samples, the other copper deposits show a uniformly high Co: Ni ratio (27–38), indicating a

sympathetic variation despite the differences in absolute amounts. At Noranda the Co: Ni ratio of the pyrites again decreases with depth (30.7, 24.36 and 14.6 in near surface, 23rd and 31st levels respectively) reflecting the fact that cobalt is decreasing more sharply than nickel with increasing depth.

Silver.—Pyrites from the copper deposits consistently show a higher silver content than pyrite from the Sudbury ores or the McIntyre gold. The variation within the copper group is slightly over a 2-fold one. No variation with depth was observed in McIntyre samples but a slight increase in silver of pyrites from the lower H ore body at Noranda (.001–.003%) is apparent at depth as noted by Auger.

Chromium.—Chromium occurs in detectable amounts in only scattered samples except at Sudbury where it is present in all samples. The mean concentration in all samples where it occurs in detectable amounts is very steady and chiefly ranges between 0.001 and 0.002 percent.

Lead.—Lead shows a very erratic distribution and is most probably due to minute inclusions of galena. No significant variation is evident between the three types of deposits except that Chibougamau pyrites are very low in lead and the Sudbury pyrites show a lower than average lead content.

Tin.—Tin occurs in detectable amounts in only four sets of pyrites from Manitouwadge, Quemont, Noranda and Normetal. The tin content of the Manitouwadge pyrites range up to 0.7 percent, but as noted earlier, much if not all of this may be due to contained cassiterite. In the other copper deposits, tin is present in some but not all of the individual sulfides and though means have been calculated they are not considered significant. It is quite possible tin is present in these sulfides as a contaminant possibly with sphalerite.

Titanium.—Titanium is relatively steady within the copper deposits, its concentration being lower than in the pyrites from the hydrothermal gold type deposit of the McIntyre and pyrite from the ores of the Sudbury area.

PYRRHOTITE

Qualitative Analysis.—In addition to elements studied quantitatively, other elements detected in pyrrhotite from the various deposits are as shown in Table 5. Gold is not readily detectable in the pyrrhotites from the Sudbury area unless concentrated by fire assay in silver beads, but is present in pyrrhotites from Flin Flon and Noranda of the hydrothermal copper deposits and present in the single pyrrhotite analyzed from the McIntyre. Arsenic is not detectable in the pyrrhotites from the Sudbury area, but is present in pyrrhotites from Flin Flon, Chibougamau and Normetal of the hydrothermal copper group studied and is also present in the pyrrhotite sample analyzed from the McIntyre. Bismuth is not found in the pyrrhotite from the Sudbury area and is lacking in the Flin Flon pyrrhotite, but occurs in the pyrrhotite from the five other copper mines and in McIntyre pyrrhotite. Manganese, molybdenum, tin, and vanadium are present in the pyrrhotites from all the deposits studied. Again there are more similarities than differences in pyrrhotites of the different deposits.

TABLE 5

QUALITATIVE ANALYSIS OF PYRRHOTITES FROM VARIOUS MINES

Location	Au	As	Bi	Mn	Mo	Sn	V
Nickel Ores							
Hardy				x	x	x	x
Fecunis Lake				x	x	x	x
Copper Ores							
Flin Flon	x	x		x	x	x	x
Chibougamau		x	x	x	x	x	x
Quemont			x	x	x	x	x
Noranda	x		x	x	x	x	x
Normetal		x	x	x	x	x	x
Manitouwadge			x	x	x	x	x
Gold Ore							
McIntyre	x	x	x	x	x	x	x

Quantitative Analyses.—The mean concentrations of the trace elements found in the pyrrhotite samples are tabulated in Table No. 6.

Cobalt.—As shown in a separate study by Hawley and Stanton the cobalt (and nickel) content of Sudbury pyrrhotites varies in deposits of the north and south ranges as was also found true of their selenium content, the higher quantities being typical of South Range deposits. Average cobalt content is for the most part above that of pyrrhotite from the copper and gold deposits but not much so, and two of the copper ore-pyrrhotites exceed the Sudbury pyrrhotites in cobalt. Pyrrhotite like the pyrite from the Campbell-Chibougamau deposit is enriched relatively in cobalt, which over a depth of 1,150 feet is relatively constant, with a slight decrease below the 900-foot level.

TABLE 6

Mean Concentration of Trace Elements in Pyrrhotites from Various Mines

Location	No. Specs.	Co%	Ni%	Co:Ni	Ag%	Cr%	Pb%	Sn%	Ti%	V%
Nickel Ores										
Misc. Mines	103	.076	1.65	.046	.00018					
Hardy Mine	10	.034	1.33	0.026	.00014	.0023	n.d.	n.d.	.031	0.0100
Fecunis Lake	14	.039	1.34	0.029	.00016	.0027	.0195	n.d.	.034	0.0094
Copper Ores										
Flin Flon	1	.15	0.019	7.9	.008	.0008	.070	n.d.	.0096	n.d.
Chibougamau	20	.11	0.032	3.4	.00091	.0015	.033	0.0023	.0080	n.d.
Quemont	3	.026	0.0034	7.6	.0016	.0021	.031	n.d.	.0091	n.d.
Noranda	9	.038	0.0016	23.7	.00085	.0015	.020	0.0071	.0074	n.d.
Normetal	8	.019	0.001	19.2	.011	.0014	.032	0.122	.032	n.d.
Manitouwadge (Geco)	9	.0072	0.0027	2.66	.0072	.0013	.037	0.10	.010	n.d.
Gold Ore										
McIntyre	1	.0550	.012	4.6	.00009	n.d.	n.d.	n.d.	.0064	n.d.

Nickel.—As might be expected the pyrrhotite samples from the copper-nickel deposits of the Sudbury type are nickel rich, due probably in part to finely intergrown pentlandite. Within the group of hydrothermal copper type deposits Normetal is low in nickel while Chibougamau is relatively high, but all are distinctly lower than the pyrrhotite of nickeliferous ores. The McIntyre pyrrhotite has a nickel content within the range of nickel concentrations shown by the copper deposits. At Chibougamau in the depth range

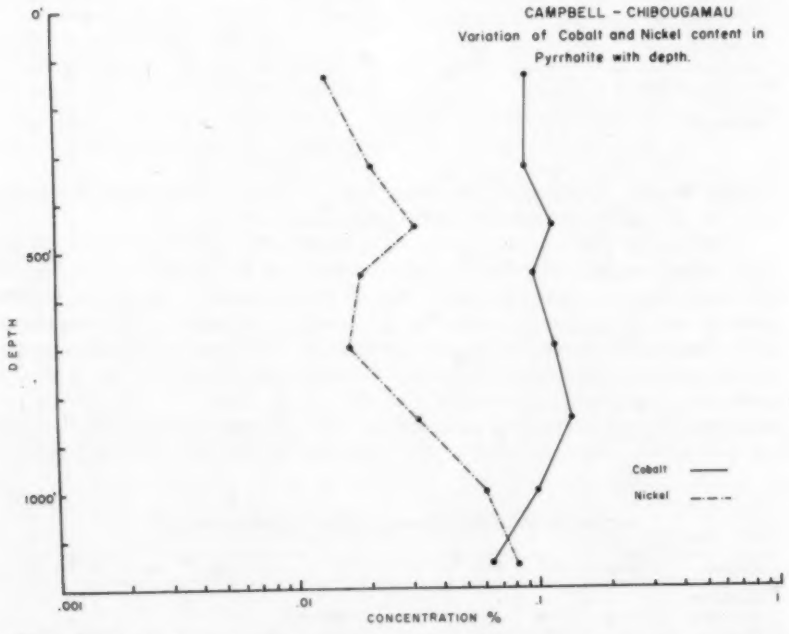

FIG. 2. Variation of cobalt and nickel content in pyrrhotite with depth. Campbell-Chibougamau deposit, Quebec.

125 to 1,150 feet nickel values show a distinct increase with depth and on the lowest level from which samples were obtained (1,150 feet) the nickel content (.082%) exceeds the cobalt content (.051%), see Figure 2.

Cobalt-Nickel Ratios.—The Co : Ni ratios of the samples from the Sudbury basin have low Co : Ni ratios due to abnormal quantities of nickel. Manitouwadge, Flin Flon, Quemont, Chibougamau and the McIntyre have Co : Ni ratios greater than one but at Noranda and Normetal the Co : Ni ratios are much higher. Although cobalt is highest in Chibougamau pyrrhotite, nickel is also relatively high, hence a Co : Ni ratio of 3.4. At Chibougamau in the depth range 125 to 1,150 feet the cobalt content remains relatively constant whereas the nickel content increases with depth; this is reflected in decreasing Co : Ni ratios with increasing depth.

Silver.—Relative to those from the hydrothermal copper deposits the pyrrhotites from the Sudbury basin are poor in silver. Within the group of hydrothermal copper deposits studied pyrrhotites from Manitouwadge, Flin Flon and Normetal carry the higher silver concentrations. The McIntyre pyrrhotite is markedly low in silver, having about one quarter the amount found in McIntyre pyrite.

Chromium.—Chromium is of erratic distribution in the copper deposits but is present in all the samples of pyrrhotite from the Sudbury basin.

Lead.—Although of somewhat erratic occurrence lead is mostly in higher concentrations in the pyrrhotites from the hydrothermal copper deposits than in the pyrrhotites from the Sudbury area. In the gold deposit it was not determined quantitatively.

TABLE 7

QUALITATIVE ANALYSIS OF CHALCOPYRITES FROM VARIOUS MINES

Location	Au	As	Bi	Mn	Mo	Sn	V
Sudbury[1]				x		x	
Flin Flon	x	x	x	x	x	x	x
Chibougamau		x	x	x	x	x	x
Quemont			x	x	(x)	x	x
Noranda	x		x	x	x	x	x
Normetal		x	x	x		x	x
Manitouwadge			x	x	x	x	x
McIntyre	x	x	x	x	x	x	x

[1] Lewis (1950) detected Co, Ni, Pb, Sn, Zn, Ag, Mn, Ti, Al, and Si in five samples of Sudbury chalcopyrites, concentrates of which also contained Pt and Pd.
(x)—present in only one sample.

Tin.—Tin is erratic in its occurrence and is restricted to the copper deposits although stannite has been recognized in certain ores of the nickel deposits, (Frood) where it is associated with sphalerite. As noted under pyrite, this same association may account for tin in pyrrhotite of several of the copper ores.

Titanium.—Titanium occurs in higher concentrations in the pyrrhotites from the Sudbury basin than in the hydrothermal copper deposits, possibly due to the fact that in ores in the quartz diorite and basic rocks, ilmenite is present and may have been replaced by sulfides with absorption of titanium. The pyrrhotite sample from the McIntyre contains slightly less titanium than pyrrhotites of the copper deposits.

Vanadium.—Vanadium though present in all samples has been determined quantitatively only in the pyrrhotites from the Sudbury basin.

CHALCOPYRITE

Qualitative Analyses.—The elements detected in chalcopyrite from the various deposits other than those determined quantitatively are shown in the Table 7.

Gold is found in detectable amounts in chalcopyrites from Flin Flon, Nor-

anda and the McIntyre. Arsenic is present in chalcopyrite samples from Flin Flon, Chibougamau, Normetal and the McIntyre. Molybdenum is present in samples of chalcopyrite from Flin Flon, Chibougamau, Quemont (single sample), Manitouwadge and the McIntyre, but is lacking in samples from Noranda and Normetal. Bismuth, manganese, tin and vanadium are present in chalcopyrites from all the copper and gold deposits.

Quantitative Analyses.—The mean concentrations of the various elements determined in chalcopyrites are shown in Table 8.

Cobalt.—In the hydrothermal copper deposits, cobalt in chalcopyrite ranges from 0.0055 to 0.049 percent, that of Chibougamau being the highest. The one sample analyzed from the McIntyre is intermediate between the two highest of the copper deposits. The cobalt content of the Sudbury chalcopyrites is 0.0154 percent, within the range of the copper deposits.

TABLE 8

Mean Concentrations of Trace Elements in Chalcopyrites from various Mines

Location	No. Specs.	Co%	Ni%	Co:Ni	Ag%	Cr%	Pb%	Sn%	Ti%	V%
Nickel Ores										
Misc. Sud.	28	0.0154	0.037	0.42	0.018	0.0079	0.019	n.d	0.0310	0.0055
Copper Ores										
Flin Flon	4	0.028	0.0013	21.5	0.016	0.0010	0.037	n.d.	0.0079	n.d.
Chibougamau	9	0.049	0.012	4.1	0.0119	0.0010	0.010	0.0016	0.0089	n.d.
Quemont	4	0.0055	0.0007	8.5	0.016	0.0009	0.012	0.0033	0.0057	n.d.
Noranda	6	0.0077	0.0014	5.5	0.0091	0.0010	0.011	0.0043	0.0073	n.d.
Normetal	8	0.016	0.0010	16.3	0.031	0.0011	0.023	0.0025	0.0230	n.d.
Manitouwadge	11	0.0088	0.0007	12.2	0.030	0.0014	0.026	0.100	0.0069	n.d.
Gold Ores										
McIntyre	1	0.036	0.0048	7.5	0.0165	0.0012	0.018	n.d.	0.0150	n.d

Nickel.—The Sudbury chalcopyrites have a higher nickel content relative to the hydrothermal copper deposits where nickel is quite low at Quemont and highest at Chibougamau. High nickel in Sudbury chalcopyrite may be due largely to fine admixtures of pentlandite incompletely separated from it. The single chalcopyrite specimen from the McIntyre has a nickel content higher than the copper deposits with the exception of Chibougamau, but well below that of Sudbury samples.

Cobalt-Nickel Ratios.—The cobalt-nickel ratio of the Sudbury chalcopyrites, due to the high nickel values noted above, are characteristically less than unity. In the case of the hydrothermal copper and gold deposits the ratios are all greater than one.

Silver.—The Sudbury chalcopyrites and that from the McIntyre mine have silver contents intermediate between the range displayed in the hydrothermal group of copper deposits which is lowest at Noranda and highest at

Manitouwadge and Normetal. No marked variation in the silver content in chalcopyrites with type of deposit is evident.

Chromium.—Chromium is erratic in its occurrence in the chalcopyrite specimens but the mean concentrations in samples where it is detectable is very constant, except at Sudbury where the concentration is considerably higher.

Lead.—The lead concentrations of the chalcopyrite specimens extend over a range of 0.10–.037, Flin Flon being high in lead and Chibougamau low. No marked differences in types of deposits are apparent.

Tin.—Tin was not determined quantitatively in chalcopyrite samples from Sudbury, Flin Flon and McIntyre, which contain only traces. Of the other deposits Manitouwadge and Normetal are both relatively high in tin. In the former some contamination with cassiterite may have occurred as noted for the other sulfides. In all chalcopyrites from Quemont (4), Noranda (8) and in 7 of 8 from Normetal, tin was detected in amounts ranging from .0013 to .011%.

Titanium.—The chalcopyrites from the Sudbury area have a relatively high titanium content. With the exception of Normetal, the McIntyre chalcopyrite sample has a somewhat higher titanium content than those from the hydrothermal copper deposits.

DISTRIBUTION OF ELEMENTS IN THE SULFIDES

With a few exceptions qualitative analyses show much the same suite of trace elements in pyrite, pyrrhotite and chalcopyrite, whether these are from magmatic nickeliferous ores, hydrothermal copper ores or pyritic gold-quartz ores. Gold clearly shows a greater preference for pyrite than pyrrhotite or chalcopyrite though quantitative data are not available. As shown in another study,† it is difficult to find pyrite in many gold deposits that is completely barren of gold. The platinum metals, as is well known, are more commonly detected in all three sulfides of nickeliferous ores, but several, particularly palladium, have also been noted as traces or better in the common sulfides of non-nickeliferous copper deposits and even gold deposits, several of which have been included in this study, and also in others.* In these, chalcopyrite commonly has the highest concentrations of Pd, Pt and Rh. As noted by the above authors data suggested the rarer metals Ru, Ir and Os are somewhat more concentrated in pyrrhotite than in chalcopyrite, but improved techniques for recovering the different platinum metals are still required before more precise data will be available according to Beamish (1959).

Distribution ratios for cobalt, nickel and silver as determined quantitatively for sulfides of the various deposits are assembled in Table 9.

Cobalt.—The distribution ratios of the cobalt concentrations in the three sulfides from the various deposits indicate that with only one exception cobalt is preferentially concentrated in the sulfides in the order pyrite, pyrrhotite and chalcopyrite. This is the usual order of crystallization in such ores, and

† See Hawley, J. E. (1952).
* See Hawley, J. E. and Rimsaite, Y., 1953.

agrees with results of Gavelin and Gabrielson. That the actual amounts and ratios may vary considerably even in one type of deposit is shown by differences in north and south range samples of Sudbury ores in which pyrite is exceptionally cobalt rich, compared with other ores. The rather uniform distribution ratios for five of the six hydrothermal copper ores are striking and seem distinctively different from that of the one gold ore studied.

TABLE 9

DISTRIBUTION OF ELEMENTS IN PYRITE, PYRRHOTITE AND CHALCOPYRITE

Location	Cobalt			Nickel		Silver	
	Mean		Distn. Ratio	Mean	Distn. Ratio	Mean	Distn. Ratio
Nickel Ores							
Sudbury	Py.	1.03		0.10		.00024	
with N. Range	Po.	0.037	27.8:1:0.41	1.33	.075:1:0.029	.00015	1.6:1:120
Pyrrhotites	Cp.	0.0154		0.037		.0180	
Sudbury	Py.	1.03		0.10		.00024	
with 103	Po.	0.076	13.5:1:0.2	1.65	0.06:1:0.02	.00018	1.3:1:100
Pyrrhotites	Cp.	0.0154		0.037		.0180	
Copper Ores							
Flin Flon	Py.	0.066		0.0057		.0066	
	Po.	0.15	0.44:1:0.086	0.019	0.30:1:0.068	.008	0.75:1:2.0
	Cpy.	0.028		0.0013		.0160	
Chibougamau	Py.	0.30		0.011		.0039	
	Po.	0.11	4.2:1:0.38	0.032	0.31:1:0.38	.00091	4.3:1:13.1
	Cpy.	0.049		0.012		.0119	
Quemont	Py.	0.084		0.0022		.0055	
	Po.	0.026	3.2:1:0.21	0.0034	0.65:1:0.19	.0016	3.4:1:10.7
	Cpy.	0.0055		0.00065		.016	
Noranda	Py.	0.116		0.0041		.0031	
	Po.	0.038	3.1:1:0.20	0.0016	2.6:1:0.88	.00085	3.6:1:10.7
	Cpy.	0.0077		0.0014		.0091	
Normetal	Py.	0.062	3.3:1:0.84	0.0016	1.6:1:1	.0042	0.38:1:2.8
	Po.	0.019		0.00099		.011	
	Cpy.	0.016		0.00098		.031	
Manitouwadge	Py.	0.031		0.0011		.0028	
(Geco)	Po.	0.0072	4.3:1:1.2	0.0027	0.41:1:0.27	.0072	0.39:1:4.2
	Cpy.	0.0088		0.00072		.030	
Gold Ore							
McIntyre (Av)	Py.	0.079		0.041		.00044	
	Po.	0.055	1.4:1:0.65	0.012	3.4:1:0.04	.00009	4.9:1:183
	Cpy.	0.036		0.0048		.0165	

Nickel.—The distribution ratios of nickel in the three sulfides are much less regular than those of cobalt. Gavelin and Gabrielson, found the order of maximum nickel concentration to be pyrrhotite, pyrite and chalcopyrite. In the present study, this is true for only four of eight deposits. In a fifth, (Chibougamau), the order is pyrrhotite, chalcopyrite and pyrite, and in Noranda, Normetal and McIntyre, pyrite is richest, followed by pyrrhotite and chalcopyrite.

Silver.—The distribution of silver between the three sulfides is shown in Table 9. In all cases chalcopyrite contains more silver than pyrite or pyrrhotite. At Sudbury, Chibougamau, Quemont, Noranda and the McIntyre, pyrite contains more silver than pyrrhotite but at Normetal and Manitouwadge the reverse is the case.

The remaining elements chromium, lead, tin, and titanium show no systematic or regular differences in distribution in the three sulfides, although they do vary in amounts with type and locality of deposit.

Lead.—Data on lead content in sulfides of the various deposits are incomplete. From Tables 4, 6 and 8, however, it is clear lead is more abundant in the sulfides of the copper ores, and in four of six deposits shows a preferential concentration in the order, pyrite, pyrrhotite, chalcopyrite. In the gold deposit more lead occurs in pyrite than in chalcopyrite, but was not detectable in pyrrhotite. Lead is obviously very low in the sulfides of the nickel ores though galena does occur in places in them.

CONCLUSIONS AND DISCUSSION

Conclusions to be drawn from the study of trace elements in pyrite, pyrrhotite and chalcopyrite formed more or less during one period of mineralization in deposits of three general types are as follows:

1. A quantitative study indicates that only in the case of cobalt, nickel, silver and lead has a specific order of preferential concentration been found in the three sulfides. This is best illustrated by cobalt and silver, but is less well defined for nickel and lead.
2. Variations in quantities of trace elements are apparent in deposits of different as well as of similar types. Variations are also evident in varieties of pyrite as at McIntyre, in pyrite of different generations and in pyrrhotites of differently located ore bodies, as at Sudbury. Variations also occur with depth in certain ore shoots.
3. Qualitatively, much the same suite of trace elements may be found in the three sulfides of all three types of deposits studied, modified as noted under (2) above.

These points will be considered in turn and possible interpretations discussed. According to the distribution ratios cobalt is with one exception, preferentially concentrated in sulfides in the order: pyrite, pyrrhotite, chalcopyrite, the normal order of their paragenesis. As might be expected since nickel can substitute for iron completely in pyrrhotite solid solutions, nickel is preferentially concentrated in the order: pyrrhotite, pyrite and chalcopyrite. In two of the eight ores studied (Noranda and McIntyre) however, the order for nickel is the same as for cobalt, but chalcopyrite is uniformly the poorest host for nickel. Silver in every case is most concentrated in chalcopyrite, at times to a very marked degree in the nickel and gold ores, but to a lesser extent in the copper ores so that in the latter more silver appears in pyrite and pyrrhotite. Lead though not detected in all samples is relatively more abundant in pyrite and pyrrhotite of the copper ores compared with the others,

but amounts in all the chalcopyrites vary rather little. From the copper ores a preferential concentration of lead in the order pyrite, pyrrhotite, chalcopyrite seems to hold.

The preferential concentration in particular sulfides is taken as a reflection that these particular elements are more acceptable in certain lattices than in others. In the case of cobalt and silver, this appears to hold, no matter what the type of ore or the absolute amounts of the metal itself. It is somewhat surprising in the case of nickel ores to find that though nickel is in excess of cobalt in both basic igneous rocks (see Table 10) with which such ores are associated, and in the ores themselves, pyrite of these ores shows an enrichment in cobalt 14 to 60 + times that in pyrrhotite and chalcopyrite respectively. This may be due in part to cobalt being more chalcophile than nickel as suggested by Wager, Vincent and Smales. Here however, we note Co is concentrated in early pyrite † whereas at Skaergaard it was found to preferentially enter the early copper rich liquid.

TABLE 10

MINOR ELEMENT CONTENT OF ROCKS

(Data from Rankama and Sahama, 1950)

Rock type	Co%	Ni%	Ag%	Cr%	Pb%	Sn%	V%
Peridotite	.0237	0.32	.0009	0.34			.0017
Gabbro	.0079	0.015	.0007	0.034	.0005	.00008	.0056
Diorite	.0032	0.004		0.0068			
Granite	.0008	0.0002	.0008	0.0002	.003	.008	.0017

Variations in quantities of trace elements with type of ore, in addition to the well known association of platinum metals with nickeliferous sulfides, include higher concentrations of chromium, vanadium and titanium, and lower silver, tin and lead in our nickel ore samples compared with others. These differences may in part be attributed to the basic rocks with which the ores are associated, and to inclusions of wallrocks within the sulfides.

Differences in composition of sulfides of the copper ores as compared with the nickel and gold ores, in general consist of more lead, in some cases tin, more evenly distributed silver, lower cobalt in pyrite and a generally lower nickel content.

Among the copper ores themselves some are clearly enriched in certain elements such as Manitouwadge and Normetal sulfides in tin, and the Campbell Chibougamau in cobalt. Noranda sulfide ores, as is well known, are relatively rich in gold compared with many other similar ores. In general when such is the case, all three sulfides show a relative enrichment compared with the sulfides of other deposits. In the same way, McIntyre sulfides are relatively richer in nickel than the copper ores but lower in silver, although the latter is a prominent constituent of associated native gold and tellurides.

Such peculiarities as these may obviously be due to either local differences

† Including pyrite formed by reaction of pyrrhotite with sulphur vapor.

in source and character of ore fluids or to contamination of similar ore fluids by differing wallrocks. In support of the latter is the occurrence of cassiterite with the sulfides in parts of the ore from the Geco deposit where many pre-ore pegmatites are located nearby. Noranda copper ores lie in a well defined gold metallogenic province, and at Chibougamau, Allard found cobalt as a minor element in associated granitic rocks.

Variations in trace elements in varieties of one mineral such as pyrite in single ore deposits as well as in widely separated ore deposits of one type are noteworthy. Varieties of pyrite which F. G. Smith distinguished by distinctly different temperature effects were shown by the senior author to vary in composition. Further study by the junior author confirms higher cobalt and nickel and lower silver in pyrite showing the higher temperature effects compared with the low. There is no other obvious explanation than differences in temperature of formation to account for such consistent differences in composition.

Variations in the cobalt content of different generations of pyrite in the Sudbury ores is probably best explained by changes in the ore fluid with time (exo-magmatic differentiation of Neumann). This may also account for the rather striking and consistent differences in the selenium content of both pentlandite and pyrrhotite from the widely separated deposits of the North Range and the South. In a separate study, not yet published (Hawley J. E. and Stanton R. L.), North Range pyrrhotites are consistently lower in cobalt, nickel and silver as well as in selenium. If the source of the ore fluids lay to the south of even the South Range, as suggested by gravity anomalies, distance of travel of ore fluids and time of precipitation for North and South Range deposits may have differed considerably and allowed some such differentiation in otherwise very similar ore fluids to occur.

Variations of the sulfides in certain trace elements with depth appear irregular. Cobalt and nickel for instance, both increase in some, decrease in others or behave in opposite manners. As the host minerals are not saturated with these elements, the content, as noted by Fryklund and Harner, can have no precise temperature significance. Causes of such variations—which, as in the case of gold in pyrite at the Kerr Addison Mine, have some economic significance—may include changes in the mineralizing fluids as they rise, and the variable character of the wallrocks undergoing replacement.

Broadly, however, what does seem of some significance is the fact that much the same suite of trace elements may be found in the three common sulfides of the three types of deposits. Where certain elements are not in detectable amounts within the sulfides, they commonly occur in other discrete minerals in the same ore. These similarities surely reflect a common or similar source, whether magmatic or metamorphic. A similar observation on the trace element content of sphalerites from deposits of different intensity was considered by Graton and Harcourt, as evidence of a common magmatic derivation, and minor variations in trace elements were recognized as being due to different conditions of deposition. Since the nickeliferous ores in this case are unquestionably magmatic, it is logical therefore, to conclude that the others are too. The efficiency of the two types of processes (magmatic

and metamorphic) in concentrating ores has been considered elsewhere, and recently Sales, has argued forcibly in favor of magmatic fluids as against a metamorphic or syngenetic origin especially for copper deposits. Recently also, Wilson and Anderson and Kilburn, in studying massive sulfide ores, in some cases from the same mines as considered here, have called attention to the similarity of metal ratios and chemical trends in the sulfide ores and in igneous rocks as supporting strongly a magmatic origin of these ores, some of which are clearly formed from sulfide liquid concentrations and others from hydrothermal fluids. This is not to say that only magmatic sources determine the final composition of such ores. Not only may exo-magmatic differentiation occur as noted by Neumann, but contamination of such fluids by any "metamorphic fluids" in the vicinity, and by solution and replacement of various wallrocks can obviously come about and is expectable. Local variations in trace elements as noted above seem best accounted for in this manner. But these are minor features and a much more rigorous demonstration of the effectiveness of metamorphic processes in ore concentration must be forthcoming before they can be considered as major factors in ore genesis.

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BRECCIA AND PEBBLE COLUMNS ASSOCIATED WITH EPIGENETIC ORE DEPOSITS

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ABSTRACT

Columns of rock fragments show a wide variety of form and structure and are associated with many different kinds of ore deposits. The fragments range from angular to rounded and from blocks weighing many tons to the particles that constitute a matrix of rock paste. The columns generally develop on planar structures or at structural intersections mostly in competent rocks.

Two broad divisions of mineralized rock fragment columns are proposed, namely pre-hydrothermal and co-hydrothermal. In the pre-hydrothermal type the spatial coincidence between the column and the ore shoots in most cases is poor because the column developed prior to and independently of mineralization, and other controls besides the configuration of the column were operative in localizing the ore. In the co-hydrothermal type there is closer coincidence in time and generally in space between ore and the column because the column results from the corrosive action of the hydrothermal fluids.

Tectonic breccias, volcanic vent fillings, intrusive igneous breccias, and filled sink and circle deposits formed by ground water leaching of carbonate rocks, generally come under the heading of pre-hydrothermal structures.

Other breccia columns, hydrothermal pebble dikes, and probably most porphyry copper deposits come under the heading of co-hydrothermal structures.

The incidence of ore most likely is higher in co-hydrothermal than in pre-hydrothermal structures though important ore deposits occur in both types.

INTRODUCTION

LIKE many of the problems in the field of ore deposits, breccia and pebble columns have long received sporadic and inconclusive consideration. The subject is broader than the name implies and the many possible modes of breccia pipe origin complicate their study. The associated rock types vary widely and so do the ore deposits. For example, such columns occur in or near the nickel deposits of Sudbury, the porphyry copper deposits of the Basin and Range Province, lead-zinc deposits in Yugoslavia, the Oruro silver-tin district of Bolivia, the uranium deposits of the Colorado Plateau, the gold-copper deposits of Queensland, the diamond deposits of Kimberly, and the gold deposits of Cripple Creek, Colorado.

This article is written with the purpose of clarifying present information on breccia and pebble columns, synthesizing this information, and calling attention to the geologic problems. Most of the material in the article comes from a review of the literature on breccia pipes and similar structures and on their relationship to ore deposits. Most of the examples cited are in the western United States though references to other areas are included. Emphasis is placed on those breccia and pebble structures that show the most direct relationship to ore deposition and which appear to be an outgrowth of hydrothermal activity. This group of breccia and pebble structures is referred to as "co-hydrothermal." Breccia and pebble structures that originated independently of hydrothermal activity, and referred to as "pre-hydrothermal," are treated separately and less completely because they comprise a much more varied group of structures that can be treated independently of ore-forming processes, as problems in structural geology and volcanology. They are important as ore solution conduits and hosts to ore deposits but do not present the same enigma as do co-hydrothermal breccia and pebble structures. Furthermore, this investigation did not include the variety of pipe-like ore deposits referred to in such terms as "manto," "massive-pipe deposit," "pipe-like pegmatite," etc. The writer believes that some of these deposits, as for example, the O. K. pipe in Utah (6, fig. 12), and pipes with quartz cores in eastern Australia (3) may have originated as breccia pipes but passed over into massive bodies by complete replacement of the original breccias.

The investigation of breccia and pebble columns was undertaken for the Southwest District of Bear Creek Mining Company and the following article is published with the company's permission. Thomas N. Walthier, Raymond F. Robinson, and others of the Southwest District proposed this subject. The writer acknowledges his indebtedness to those who read his report and offered suggestions for its improvement, particularly John W. Vanderwilt, Truman H. Kuhn, Rudy C. Epis, William R. Jones, Frank Simons, and

Donald G. Bryant. In its present form the report does not necessarily reflect the views of these critics.

TERMINOLOGY

The term "breccia pipe" as used in the geologic literature refers in most cases, to a crudely cylindrical, highly inclined or vertical structural unit composed wholly or partly of angular, or rounded, rock fragments with or without a matrix. "Breccia chimney" is a synonymous but less popular term. Though "breccia pipe" is not broad enough to be descriptive of the wide variety of form and structure that characterizes these bodies, the term is retained here because of its well established use. In this report it is used in a broad enough sense to include forms widely divergent from that of a pipe and to include rounder constituents than implied by the word "breccia".

In describing the intrusive breccias at Toquepala, Peru, Richard and Courtright (44, p. 264) state: "restrict the term *breccia* to formations composed *entirely* of fragmental material in which the fragments have been rotated and displaced, in contrast to rock that has been merely intricately jointed and, also, to rock that was emplaced in a fused condition." They apply the term agglomerate to an intrusive breccia having a fused matrix. This is not appropriate in view of the extrusive connotation usually attached to "agglomerate." In the present report "breccia pipe" is considered to include breccias either with or without a fused matrix. The pipes composed of "intricately jointed" rock have been referred to by Lacy (31, p. 105) as "shatter pipes" and this term is adopted here. According to Bateman (2, p. 133), Currier distinguishes between "rubble" breccias, and "shatter or crackle breccias" among the Appalachian zinc deposits. This is essentially the same distinction as between breccia pipes and shatter pipes. Lacy (31) applies "breccia pipe" to a pipe-like body of brecciated rock in which the rock fragments have moved generally upward and have been rotated." This stipulation as to movement of rock fragments is too restrictive for general use. In most references to breccia pipes, like the last one mentioned, there is a genetic connotation, one of the most common being that they are "explosion breccias." There are so many possible origins for breccia pipes that the term by itself should not imply any specific origin.

The term "diatreme" is, in many cases, used as an equivalent to "breccia pipe." However, a diatreme (22, p. 80) is "a pipe or vent drilled through enclosing rocks (usually flat-lying sedimentary rocks) by the explosive energy of gas-charged magmas," whereas in many pipes associated with ore deposits there is little evidence for such a dynamic origin.

"Pebble dikes" connected with igneous or hydrothermal activity are tabular bodies in which the fragments are rounded due to a milling action and/or corrosion by gaseous or aqueous fluids. Their origin, together with other characteristics to be described later, distinguishes them from the sedimentary bodies known as "clastic dikes." In the AGI Glossary (22, p. 52, 214) no genetic significance is attached to the term "pebble dike." It is defined simply as a clastic dike "composed largely of pebbles." The pebble dikes described in this report are the hydrothermal, not the sedimentary type.

"Breccia dikes" are tabular, but unlike pebble dikes, the fragments may be angular as well as rounded and the matrix fused.

"Breccia reef" usually refers to a silicified fault breccia.

Two terms introduced in this article are "co-hydrothermal" and "pre-hydrothermal." *Co-hydrothermal* breccia and pebble columns were formed by "mineralization stoping" which was at first localized by some structure such as a fault intersection, an intrusive contact, or a zone of tectonic breccia. In this type, brecciation occurs mainly because of collapse triggered by volume reduction due to the corrosive action of hydrothermal fluids. The term co-hydrothermal is applied because brecciation was more or less contemporaneous and coextensive with mineralization and an effect of the mineralizing fluids themselves. A volcanic environment commonly is absent or, if present, shows no spatial relationship with the pipes. Most porphyry copper deposits probably belong in this category and some massive, pipe-like replacement deposits may have originated as co-hydrothermal breccia pipes.

Pre-hydrothermal breccia and pebble columns are bodies of locally mineralized fragmental rock formed before and independently of mineralization. In the pre-existing breccia host that was favorable to the transmission of ore solutions additional controls such as post-breccia structures or a subjacent cupola acted to localize ore deposition. In this type there is distinctly less spatial coincidence between mineralization and the breccia body. Although most pre-hydrothermal breccia pipes are volcanic structures, such as vents or calderas, breccias formed in other ways also localize ore. Pre- and co-hydrothermal structures overlap because hydrothermal activity can further brecciate a pre-existing breccia.

GENERAL FEATURES OF BRECCIA AND PEBBLE COLUMNS

Form and Dimensions.—Breccia and pebble columns show a wide variety of form. They may be dike-like, pipe-shaped, shaped like an inverted cone, or be so irregular as to defy verbal description. Many of them taper with depth and become dike-shaped. Those roughly circular or elliptical in plan commonly are several hundred to several thousand feet across, and most of them are several times as deep as they are wide. Their vertical extent is difficult to appraise, not only because of erosion but also because relatively few of them have been bottomed by mining operations.

Of the two main types of structure that control ore deposition, as distinguished by Butler (7, p. 3), namely, "shear breaks and fracture or brecciation breaks," breccia and pebble columns would coincide with the nearer-surface, "brecciation breaks." Other features that in many cases suggest moderate to shallow, rather than great depth of origin, are (1) relative scarcity of hypothermal mineralization, compared to mesothermal and (2) distinct vertical zoning of minerals or metals.

Rock Fragments.—The constituents of breccia and pebble columns range from angular to well-rounded and from minute particles to blocks weighing many tons. Many of the fragments are hydrothermally altered on the surface and show evidence of corrosion.

Regarding movement of fragments in mineralized pipes, Joralemon states, (26, p. 255): "In most of the hundred or more chimneys I have seen the fragments are either of the adjoining wall rock or of a rock that forms a wall of a chimney at a slightly different elevation—usually higher."

Matrix.—The matrix is a rock paste composed essentially of the same material as the fragments, or an igneous rock representing a magma that has solidified around the fragments. Where silicification of the rock paste forms a firm vitreous-looking material it may be difficult to distinguish between this cemented rock paste and an igneous matrix.

Some pipes have open cavities which contain mineral druses. In others the matrix and fragments are replaced and cemented by minerals.

Relationship to Other Geologic Features.—Breccia and pebble columns generally form in competent rocks and rocks that do not have outstanding directional properties. In many cases pipes show a system of spiral fractures around their axis, for example, at Kidston, Australia (13, p. 4) in the Shoshone Range, Nevada, (18, p. 807), and at Red Mountain, Colorado (5, p. 175). They are controlled as well by faults, fissures, contacts, and structural intersections. Some pipes develop in the zone of arcuate breaks that borders calderas, as along the edge of the Silverton Caldera, Colorado (5, p. 151). Some, like the Whim Hill breccia (27, p. 289) in the granodiorite porphyry stock at Santa Rita, New Mexico, have no evident structural control.

Many pipes are by-products of hypabyssal igneous activity, and are apically situated with respect to the intrusive. W. H. Emmons (13, p. 14) speculates that breccia pipes develop above conical cupolas whereas parallel veins develop above elongated cupolas. At Bagdad, Arizona (1, pl. 3) the breccia pipes overlie projections of elongate outcrops of quartz monzonite that radiate like fingers from the exposed mass of quartz monzonite, and at Copper Basin near Prescott, Arizona (25, pl. X) most of the pipes are located near the south extremity of an elongated body of quartz monzonite porphyry, especially near its contact with the rocks it intrudes.

Examples of breccia columns situated in the intrusive itself are found at Santa Rita, and at Bisbee, Arizona.

As a general rule co-hydrothermal breccia and pebble columns are more or less contemporaneous with the close of the magmatic cycle and the beginning of hydrothermal activity.

Breccia and pebble columns are like many other structures that serve to localize ore—that is, only a small percentage of these structures carry ore and a relatively small part of the structure localizes most of the ore. For breccia and pebble columns as a whole, Walker's estimate (50) that only one percent are mineralized seems reasonable because many of these structures are tectonic or volcanic phenomena and have no direct genetic relationship to ore deposits. On the other hand, where these structures occur in mining districts, they are likely to be guides to ore either because they indicate a buried cupola or else, because they serve as loci for the transmission or accumulation of ore solutions, as at Cananea, Mexico (41, p. 703). It appears likely that the co-hydrothermal type of breccia or pebble column, being di-

rectly connected with hydrothermal processes, shows a higher incidence of mineralization than do other types.

PRE-HYDROTHERMAL BRECCIA AND PEBBLE COLUMNS

Tectonic Types.—Breccias due to faulting and folding are numerous. They usually can be recognized as tectonic by their intimate association with other tectonic features. Their importance in ore deposition is well known.

Distension of the crust due to doming (Noble, 39; Wisser, 54) might be accompanied by pipe-like collapse structures but the writer found only one account of this in the literature (Paige, 40, p. 46) and this refers to some of the ore deposits at Tyrone, New Mexico. Tectonic breccias, however, probably provide the rock fragments later incorporated in many other breccias.

Igneous Effusive Columns.—Some mineralized columns of rock fragments are partly the result of volcanic eruption. The Cresson "blowout," Colorado, probably is of this type. Mineralized volcanic vents generally belong to the class of volcanoes known as diatremes. But even diatremes fall short of the explosive violence frequently attributed to breccia pipes. Diatremes probably did not emit much, if any, lava, and they were relatively short-lived.

According to Howel Williams (52, p. 315) "accidental fragments found in most diatremes indicate depths of origin between 2,000 and 10,000 feet although some fragments in the diamond pipes of South Africa must have risen from very much greater depths." In some of the Navajo-Hopi diatremes, Arizona, Williams believes (51, p. 171) the fragments came from less than 3,000 feet below the surface.

In the Navajo-Hopi diatremes Shoemaker (47, p. 180, 182) observed that fragments larger than 10 feet in diameter generally are found at some level below their normal stratigraphic position and their net downward displacement is as great as 2,000 feet. Smaller fragments, on the other hand, were displaced either up or down. Many of these diatremes are filled with bedded tuff, some of which is water-lain. These tuffs commonly dip gently toward the center of the vent, steepen with depth, and "are commonly vertical or intensely deformed in the deeply eroded vents" (47, p. 180).

Rust (45, p. 61) in studying the diatremes in southeast Missouri, found fragments of Devonian fossils at levels 3,000 feet below their source beds. He also found that the peripheries of many of the diatremes were intensely shattered and he explained this by "counter explosion" (p. 73), due to gas pressures being built up in the pores of the wall rock surrounding the tube and, upon explosive eruption, a correspondingly sudden expansion of these pore gases toward the open tube, shattering the tube walls. Possibly such a process helped to prepare the ground for mineralization in some pipes.

The upper parts of diatremes, where they are relatively uneroded flare outward and the central portions are cylindrical. The Navajo-Hopi diatremes, according to Hack (19, p. 269) "appear to narrow downward and may even merge with a system of narrow fractures."

The diatremes of southwest Missouri are believed by Rust (45, p. 70) to represent finger-like upward extensions from the hanging walls of dikes.

Such a relationship might explain the apparently random areal arrangement of some diatremes; where the diatremes connect down-dip with a dike there might be no apparent spatial relationship to this dike at the surface. Daly (11, p. 391) believed that diatremes consistently overlie sills or laccoliths, but Hack (19, p. 370) presents evidence to show that the Navajo-Hopi diatremes probably are not underlain by a laccolith. These diatremes occupy a shallow structural basin where there is no evidence of doming.

In some diatremes the pressures may have come from an advancing magma column. In many of them, deposits of tuff indicate explosive volcanism as a driving force. Williams (52, p. 316) describes the formation of diatremes as follows: "Rising gases 'lubricate' the roof rocks; pulsating magma brecciates the cover; repeated explosions comminute the cap rocks so that frothing magma is intimately mingled with them and, near surface, steam eruptions caused by heating of ground water facilitate the drilling process while slumping of slabs from the conduit walls enlarges the conduits."

Regarding the pebble-breccia body at Toquepala, Richard and Courtright (44, p. 265) say: "It would seem to have been produced by a milling action in which the rock fragments were actively suspended in an actively circulating medium such as water or gas. In this state this pebble breccia may have had an intrusive mobility, although portions of it may also have been formed more or less in place."

The process known in industry as "fluidization" (transportation of solid or liquid particles by gas flow) and proposed by Reynolds (43) as a geologic process might very well provide the churning action needed to explain the above-mentioned pebble breccia and pebble dikes in general. Reynolds states: "Experiments have shown for example that the agitation and turnover of solid particles by water is insignificant in comparison with the rapid rate of circulation and mixing in the turbulent expanded bed of a solid-gas system" (p. 578). Fluidization conceivably could cause upward transportation, or selective transportation to allow subsidence of larger fragments and ascension of smaller ones—a process that took place in some diatremes and pebble dikes. By comparison with industrial fluidization, Reynolds believes gas transportation can be recognized by: "The association of turbulent flow structure with abraded and well-rounded rock fragments which have not apparently been transported away from their source rocks, together with lack of grading of the fragments concerned, and the possible presence of druses" (p. 579). Except for a lack of druses, the breccia bodies at Sudbury and pebble dikes in general fit these specifications quite well.

Some of the Red Mountain, Colorado, breccia pipes quite possibly developed through fluidization, judging from Burbank's description of their textures (5, p. 174): "Textures suggestive of flow structure or squeezing in the matrix of the breccia, together with mineralogical changes, suggest that in the final stage of its formation the breccia was saturated with volcanic gases or vapors and reacted to forces by plastic flow rather than by fracturing." This was written before Reynolds had proposed the fluidization theory.

It is reasonable to suppose also that water as well as gas might be a medium of transportation for intrusive breccias. In parts of some pebble

dikes (?) at Butte, stratification is evident (46, p. 20). At the Stantrg lead-zinc breccia pipe in Yugoslavia the gangue minerals are mainly quartz and carbonates, and at the surface, partly obscuring the pipe, is a travertine deposit. "In the neighborhood of the ore body feeders of water aggregating 1,000 gallons per minute issue from cavities and fissures at 85° F" (17, p.

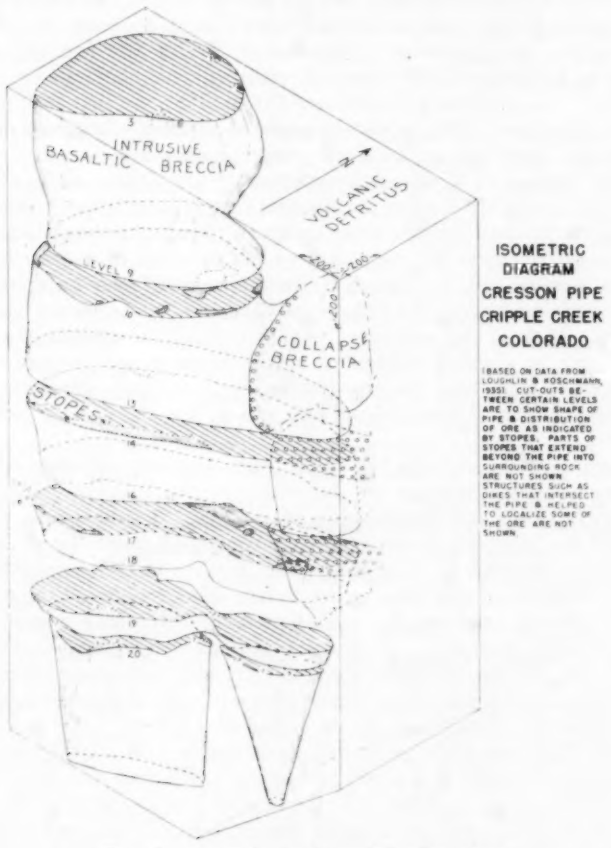

INTRUSIVE
BASALTIC BRECCIA
VOLCANIC DETRITUS
LEVEL 9
COLLAPSE BRECCIA
STOPES
ISOMETRIC DIAGRAM
CRESSON PIPE
CRIPPLE CREEK
COLORADO
BASED ON DATA FROM LOUGHLIN & KOSCHMANN, 1935. CUT-OUTS BETWEEN CERTAIN LEVELS ARE TO SHOW SHAPE OF PIPE & DISTRIBUTION OF ORE AS INDICATED BY STOPES. PARTS OF STOPES THAT EXTEND BEYOND THE PIPE INTO SURROUNDING ROCK ARE NOT SHOWN. STRUCTURES SUCH AS DIKES THAT INTERSECT THE PIPE & HELPED TO LOCALIZE SOME OF THE ORE ARE NOT SHOWN.

FIG. 1. A pre-hydrothermal breccia pipe.

300); a higher temperature of outflow than in other parts of the district. The unusual concentrations of phosphates in some of the diatremes of the Hopi Country is suggestive of thermal solutions (47, p. 180).

The above examples of deposition by water imply a rather quiet welling action, but the question should be raised as to whether the criteria advanced by Reynolds for recognition of textures resulting from fluidization would be substantially different from those for textures due to turbulent water flow

such as might occur in a boiling spring. This question is suggested by the fact that Peters has proposed such an origin for the Bull Domingo Pipe, Colorado (42).

The Cresson pipe, Cripple Creek, Colorado, is a breccia column that appears also to be a diatreme. It shows less evidence for explosion and collapse than do most diatremes, and the breccia matrix appears to have been at least partly fused. It is shown in Figure 1 as an example of a pre-hydrothermal pipe. The pipe has a diameter of 500 to 800 feet and has been followed to a depth of 2,200 feet. It changes with depth to two root-like extensions. Loughlin and Koschmann (34, p. 255) say that the "roots have evidently developed along a fissure zone of eastnortheast trend and steep southerly dip, and they are localized along the intersection of that zone with dike-filled fissures of north-northwest trend." According to Lovering and Goddard (35, p. 293) the basaltic breccia that fills the pipe "contains fragments of breccia of all the dike rocks in the surrounding ground, besides many fragments of dark-gray or greenish gray dense basalt in a dense matrix of basalt and basaltic tuff." Though the breccia resulted from an upward surge of the underlying magma the process perhaps was not as violent as in many diatremes. Most of the explosive breccias that fill the caldera in which the Cresson pipe is situated originated at the beginning of volcanic activity, whereas, the Cresson pipe developed at the end of volcanic activity after many intervening periods of intrusive activity and as the last episode before mineralization.

Though the great majority of volcanic vents carry no important mineralization, some contain important ore deposits. At the Cresson "blowout," according to Loughlin and Koschmann (34, p. 316): "A study of the ore bodies shows that the positions of the largest are controlled mainly by the junctions of the rim of the blowout with mineralized fissures in the surrounding rock, especially those along and intersecting certain dikes. Some of the ore bodies within the blowout are controlled mainly by fissures that border large included masses of latite phonolite, . . . but are also connected with fissures along the rim." In the Bassick Mine, Colorado, the gold and silver mineralization is confined to a narrow pipe-like zone within a column of andesite agglomerate having horizontal dimensions on the order of 1,200 to 1,500 feet (10, pl. XXXIV). The ore body, according to Emmons (12, p. 431) is elliptical with diameters from 20 to 100 feet and has been followed to a depth of 1,400 feet. Emmons states (p. 438): "At the time of (ore) deposition the agglomerate in what is now the ore channel was probably in no (way) essentially different from that in which the country rock is now, except that it may have been somewhat shattered along the fracture planes that are supposed to have determined the course of the ore-bearing currents." In other words, the mineralization seems to have been controlled by fractures cutting the agglomerate rather than by the configuration of the agglomerate column.

In both of these examples the agglomerate and breccia is a favorable host rock only where additional controls help to localize ore deposition.

Igneous Non-effusive Columns.—A magma, in the process of intrusion, may entrain enough solid material to form a breccia. This material can be de-

rived by magmatic stoping, by abrasion of wall rocks, if the magma is viscous, or by the engulfment of other breccias. An intrusive breccia also can originate where a chilled border or front is brecciated and then inundated by a surge of underlying magma. Sills of prophyry near Pando, Colorado, were accompanied by an "advance guard" of intrusion breccia, according to Tweto (49, p. 527).

It is likely that intrusive breccias generally follow faults and fissures and therefore are less pipe-like in form than other breccia or pebble bodies. In some places, however, they fill vents or other pre-existing pipe-like structures.

A magma also might form a breccia column by withdrawal, causing collapse of the overlying rocks. This explanation has been advanced for calderas, and some breccia columns show a remarkable structural similarity to calderas. The Pilares pipe in Sonora, Mexico, is one of these. The breccia body there is oval in plan, 1,000 to 2,000 feet across, and is developed to a depth of 1,800 feet. "The latite-andesite contact 300 to 500 feet deep outside the pipe, sags inside as an inverted cone to the 900-foot level and shows over the entire oval an average subsidence of 200 feet" (33, p. 436). Locke compares the "crackling" of this subsided body to that at Bingham Canyon, Utah, "though a little coarser" (p. 433). The periphery of the body is composed of a sleeve of jumbled breccia in a zone about 50 feet thick. Most of the ore lies in this breccia zone. Locke believed that the subsidence at Pilares was due to a process which he named "mineralization stoping," to be described later in this article. An alternative explanation would be that of magma subsidence, embodied in D. D. Smythe's explanation (48) of the pipe. Smythe postulates a pre-subsidence episode of doming due to magma intrusion, because in the lower levels of the pipe there is a faulted wedge of dacite tuff 100 feet above its normal stratigraphic position outside of and adjacent to the pipe.

Filled Sink-Structures and Circle Deposits.—Bretz (4) has described pipe-like deposits in Missouri caused by groundwater solution of limestone and dolomite. Filled sinks are funnel- or bowl-shaped mass-subsidence structures formed at depth when subsidence keeps pace with solutional removal of underlying carbonate rock. Circle deposits are cone- or bell-shaped bodies of breccia formed by progressive roof-collapse of solution caves. Mineable deposits of clay, coal, pyrite, hematite, lead and zinc, occur in some of these structures. A discussion of these mineral deposits is beyond the scope of this report. However, collapsed beds in diatremes have a counterpart in the filled-sinks, and some breccia bodies, as at Bisbee, have a counterpart in the circle-deposits. These similarities suggest that subsidence and collapse result from igneous or hydrothermal as well as from sedimentary processes.

CO-HYDROTHERMAL BRECCIA AND PEBBLE COLUMNS

Many breccia and pebble columns appear to be so closely connected to hydrothermal activity, both in space and in time, that they naturally can be termed "co-hydrothermal."

In some cases the distinction between a co-hydrothermal and a prehydrothermal structure may be difficult or impossible to make, and the two types probably overlap to some extent. For example, a collapse structure like the Pilares pipe might be due either to magma subsidence or to the corrosive action of hydrothermal fluids or to both of these processes. There are enough pipe structures, however, that are clearly related to hydrothermal activity to justify this special category.

Pebble Dikes.—As Cook (8, p. 76) remarks of the Tintic District, "The close temporal and spatial association of the monzonite intrusives, pebble dikes, hydrothermal alteration and metallization suggests a common origin." At the Plutus Mine in Tintic (15, p. 362) pebble dikes and bodies of ore are contiguous in the same fissures. However, the relationship between pebble dikes and ore commonly is not so direct. Billingsley (20) believed that pebble dikes are an indication that fissures persisted to depth and tapped igneous emanations. At Bisbee, pyrite pebbles in the pebble dikes and a lack of any fragments of ore, indicate that the dikes formed after the early surge of pyrite mineralization but before copper mineralization (personal communication, Donald G. Bryant, July 8, 1959). Nevertheless, at Cananea, according to Perry (41), pebble dikes follow post-mineralization fissures and cross-cut bands and zones of mineralization. Where pebble dikes preceded the period of ore deposition their lack of mineralization probably is explained by their relative impermeability. At Tintic, according to Kildale (28, p. 109), "they were frequently tightly cemented by the earliest solutions that carried only silica and pyrite and were thus sealed off from the later solutions carrying most of the valuable metals." Also, at Tintic, some pebble dikes are traced downward into monzonite porphyry dikes (38, p. 38), and such dikes might have sealed off the ore solutions.

Many descriptions of hydrothermal pebble dikes compare their appearance to that of concrete aggregate. The fragments in pebble dikes are, in most places, well rounded. At Tintic the pebbles have "onion structure and (15, p. 370) these pebbles are found as much as 6,000 feet above their source as well as adjacent to their source. Some of the breccias at Bisbee are in late Paleozoic rocks and contain pebbles of Precambrian Pinal Schist, indicating an upward displacement of approximately 5,000 feet (personal communication, Harry E. Metz, Nov. 29, 1956).

A detailed study of structures similar to the above pebble dikes was made at Sudbury, Ontario, by Fairbairn and Robson (14). Though called "breccias," the fragments there are subangular to rounded. Some fragments were transported thousands of feet and some are quite local in origin.

Quoting the writers (p. 1).

> Aqueous fluids at elevated temperature and pressure permeated strikewise fractures along which there was slight dislocation of the walls. Fragments were torn loose and comminution was aided by action of the fluids. This formed a mobile matrix which in time attained sufficient volume to transport the remaining fragments. Low confining pressure permitted easy movement of the breccia and prevented the development of good crystallographic orientation in the minerals. The magmatic fluids enriched the matrix particularly in sodium and in general increased the

amount of mica, chlorite sodic plagioclase, etc., over that found in the fragments . . . following these changes came the introduction of the nickeliferous sulphides.

The process of fluidization described under "Igneous Effusive Columns" seems the most plausible general explanation for hydrothermal pebble dikes. Like diatremes, pebble dikes contain fragments that traveled considerable distances, as well as fragments that are fairly close to their place of origin. The lack of a fused matrix in pebble dikes indicates an aqueous or gaseous medium of transportation for the pebbles, and the rounding of the pebble shows that attrition and possibly chemical corrosion took place. The dikes at Tintic, Bisbee, and Sudbury appear to have formed during an early surge of hydrothermal fluids, and represent a sludge formed by the flushing out of fracture and fault fillings.

The resemblance of pebble-dike fillings to that of some vents, like the "pebble breccia" at Toquepala for instance (44), and the Braden Formation (21, p. 874), suggest that among co-hydrothermal bodies, these dikes represent the closest approach to volcanic activity, and can perhaps be considered as a link between pre-hydrothermal and co-hydrothermal.

Breccia and Shatter Pipes.—Some breccia and shatter columns in or near plutons show no evidence of volcanic origin, much less, of explosive activity. Some of these pipes contain ore deposits or are associated with ore deposits. They are numerous enough compared to mineralized volcanic vents to justify McKinstry's remark, (37), "the role of volcanic explosion in creating ore-bearing pipes seems much less important than an uncritical reading of the literature might suggest."

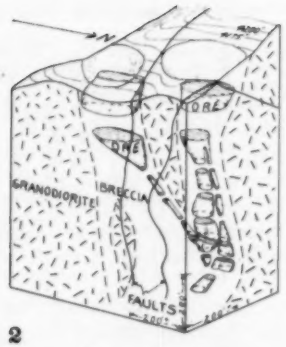
Some of these pipes are intrusion breccias, but most of them show evidence of collapse and do not contain a fused matrix like most intrusion breccias.

Locke (33, figs. 14, 15), pictures the ideal breccia pipe as carrot shaped, with the slender end down. He says (p. 448) that the Southwest and the Catalina pipes at Bisbee show a "rapid convergence of the walls forming a rude keel below which the silicified breccia descends only in irregular protuberances. . . . The pipes merely end downward as they do horizontally against fresher, less fractured country rock."

Most breccia pipes are not uniformly brecciated. Three concentric zones of greater or lesser brecciation often can be recognized. These are (25, plate VIII):

1. A core of well-brecciated rock where the porosity is relatively high, and rounded fragments abundant.
2. A less-brecciated sleeve surrounding the core, with relatively low porosity and little or no rounding of fragments.
3. A transition zone to normally fractured country rock. In zones 1 and 2 there has been rotation and some mixing of fragments, but zone 3 has the character of a shatter pipe. In other words, many column-like structures are combination breccia and shatter pipes.

The Childs-Aldwinkle copper-molybdenum-bearing pipe, Copper Creek, Arizona, shown in Figure 2, has been described by Kuhn (29, p. 518, 520,



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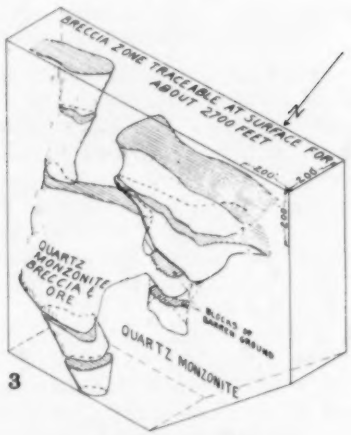
BLOCK DIAGRAM
CHILDS-ALDWINKLE ORE DEPOSIT, ARIZONA

(BASED ON DATA FROM KUHN, 1938 PLATE XXXIX, B, 1941 FIG. 71. GROUND BETWEEN ALTERNATE LEVELS OMITTED TO SHOW SHAPE OF ORE PIPES. CONTACTS BETWEEN BRECCIA & GRANODIORITE SHOWN ONLY BY TRACE ON BLOCK SURFACE, HIGHLY SCHEMATIC. FAULTS ALSO SHOWN ONLY BY TRACE ON BLOCK SURFACE. BRECCIA FRAGMENTS ARE GRANODIORITE. THE ORE IS MINERALIZED BRECCIA.



3

ISOMETRIC DIAGRAM
CACTUS ORE DEPOSIT, UTAH

(BASED ON DATA FROM BUTLER, 1913, PLATE XXIII. CUT-OUTS AT VARIOUS LEVELS ARE TO SHOW SHAPE OF DEPOSIT. FAULTS ARE NOT SHOWN, BUT THEY HAVE SMALL DISPLACEMENTS & HAVE NO DOMINANT DIRECTION.

FIG. 2. A co-hydrothermal breccia pipe.
FIG. 3. Possible co-hydrothermal porphyry copper structure.

521). This description is quoted at some length because the writer believes that this pipe is a good example of the co-hydrothermal type. The pipe:

consists of angular blocks of altered granodiorite cemented by gangue and ore minerals. In places semi-rounded fragments of granodiorite show alteration envelopes paralleling the surface of the block. . . . In the central portion of the breccia mass on the 820-foot level, the granodiorite blocks are entirely replaced by sericite, quartz, and sulphides. . . . The outline of the blocks can be seen. The zone surrounding the central core is composed chiefly of chlorite, quartz, and orthoclase with minor apatite and sulphides. . . . The cementing material also changes from principally chlorite and quartz at the bottom of the mine to sericite and quartz in the upper levels. At the borders, the breccia usually grades into soft, much altered granodiorite, and this, in turn, into hard jointed granodiorite.

Exploration by winzes and diamond drilling strongly suggests that the area of brecciation has diminished greatly at 900 feet below the surface.

Three N.10°W. faults at the bottom of the mine suggest a fracture zone. . . . Minor displacements have occurred after the formation of the breccia. . . . All major faults contain sulphides, indicating pre-mineral origin.

Regarding origin, Kuhn (29, p. 526) says:

At Copper Creek the pipes are in strong fracture zones or at intersections of fractures. Solutions that rose along these fractures altered and dissolved the rocks and thus produced breccia. Further brecciation may have resulted from slumping, similar to that described by Locke in the Pilares mine. All gradations exist between pipes intensely altered and brecciated to those that show a strong fracture pattern with very little alteration.

Another example of a co-hydrothermal pipe is the pyrite-chalcopyrite-hematite-bearing Cactus Pipe, Utah, shown in Figure 3. The ore bodies seem to have been coextensive, in great part, with the breccia except at the surface where the brecciated zone, without widening in one dimension, flares in the other dimension to a length of about 2,700 feet (6, p. 174), but the ore shoot within this zone extends along it for only about 700 feet. According to Butler (p. 175): "the limiting of the ore body is doubtless due to two influences—the massive character of the monzonite, which prevented free circulation of solutions in the early stages of mineralization, and subsequent to the faulting the gangue has still further prevented circulation. . . ."

Butler (6, p. 175) believed that brecciation in the Cactus Pipe was "due to slight movements resulting from readjustments in the cooling monzonite body." Locke's idea of mineralization stoping had not yet been proposed.

Though mineralization in co-hydrothermal breccia and shatter pipes shows a closer coincidence with the pipe boundaries than it does in the pre-hydrothermal pipes, there are pipes that probably are co-hydrothermal, and yet contain virtually no ore.

The breccia pipes near the disseminated copper mine at Bingham, Utah, are an example of this. Walker (50) speculates that some pipes may act as ore-fluid conduits but do not contain ore because temperatures in the pipe were too high for ore deposition during the period of pebble accumulation; but pipes such as the ones at Bingham are at approximately the same level as the ore deposits though off to one side of the center of porphyry copper mineralization. It seems logical to assume that such pipes are barren simply because they are not close enough to the main channels of ore deposition. Some pipes may of course simply be post-ore-fluid, but generally this is not the case.

Another explanation for barren pipes near ore is contained in Burbank's description (5, p. 183) of the volcanic pipes at Red Mountain, Colorado, where ore bodies are found within the pipes and also exterior to the pipes. The pipes not plugged by porphyry generally contain ore, whereas, the plugged pipes generally do not, although ore may occur in the surrounding ground. This same principle, namely, the sealing off of breccia by magma may apply elsewhere.

While it appears that co-hydrothermal breccia and pebble columns develop toward the close of magmatic activity and the beginning of hydrothermal activity, these two periods commonly overlap and therefore breccia pipes in any locality may show a variety of time relationships to the associated rocks. The breccia pipes at Bingham, according to Hunt (24, p. 53) are "earlier than the last mineralization but later than much of the monzonite." At Bagdad, Arizona, (1) breccia pipes range in age from older than the earliest Cretaceous (?) or Tertiary (?) volcanics to younger than the youngest igneous rocks of these periods of igneous activity. Some are mineralized and some not. Some of these pipes, however, probably are co-hydrothermal but others possibly are not.

In pebble dikes intrusive activity seems to predominate, but in co-hydrothermal breccia and shatter pipes, collapse, whether it causes great displace-

ment or hardly any displacement, is a more important process. The evidence for this is in the prevalence of a net downward movement of fragments in the co-hydrothermal breccia pipes, and in-place, unrotated fragments in the shatter pipes. With examples from Bisbee, Wisser (53, p. 772) has shown that failure of rocks above mine openings and above oxidized ore bodies assumes the form of a cylindrical, brecciated mass, except where collapse occurs relatively close to the surface, in which case the ground fails as a cylindrical unit. In mining operations, caving sometimes produces vertical, cylindrical breccia masses.

Locke (33) proposed "mineralization stoping" to explain collapse breccias in mineralized pipes. He describes this process as: "brecciation due to removal of support by the dissolving action of solutions during an early stage of their activity followed by mineral deposition in the brecciated mass at a later stage" (p. 431). The term "mineralization stoping" as understood here from Locke's paper means that there was enough corrosive action along fractures to trigger collapse of a block of ground, and this collapse was the main cause of brecciation. Once an opening had occurred collapse might be augmented by rock bursts (18, p. 809–812). Locke's theory explains both the brecciation and the ore deposition by a single process, and has in its favor the widespread evidence of corrosive action and replacement in breccia pipe deposits.

The Porphyry Copper Deposits.—Locke (33) extended the concept of mineralization stoping and breccia pipes to the porphyry copper deposits. He writes as follows (p. 452):

> The irregular disseminated mineral deposits have in no case had their shrinkage measured or its cause determined, yet their brecciation and that of the pipes are so closely analogous as to indicate that such deposits in general, including those of Bingham, Braden, and Chuquicamata, will be established as cases of the same process, confused by the overlapping of subsidence circles.
>
> The same process gives promise of application also to the tin-bearing and tourmaline-bearing stockworks, and it may even prove in some places to have been a preparatory stage for igneous intrusion.

This plausible explanation for porphyry copper shattering seems to have had little or no impact on the theory of the origin of these deposits.

Shrinkage of cooling igneous bodies has been proposed by Hulin (23) as an explanation for collapse and attendent brecciation. Against this theory is the lack of spatial coincidence between porphyry copper intrusives and the associated shattering and also the dearth of porphyry-copper-type shattering in unmineralized hypabyssal intrusives.

Possibly Locke's ideas have been interpreted as violating Lindgren's "law of equal volumes" (32, p. 92). This "law" does have exceptions, as shown by Cooper in his studies near Johnson Camp, Arizona, where stratigraphic measurements of the carbonate rocks show that "the silicated facies is as much as 30 percent thinner than the unsilicated facies" (9, p. 577). Furthermore, volume-for-volume, replacement as advocated by Lindgren, was meant primarily, it seems to the writer, to explain the lack of such *increases* in volume as might be expected from chemical reactions under atmospheric

pressure rather than any chemically indicated *decreases*. Lindgren wrote (p. 92):

> It is necessary to distinguish between (1) metasomatic changes proceeding in free crystals or grains, or in loose or plastic aggregates under light load, where force of crystallization can easily overcome the restraining pressure; and (2) metasomatic changes proceeding in rigid rocks where the new mineral is forced to make room for itself by solution of the host mineral.
>
> In the first case the volume changes proceed according to the chemical formula.

Though Lindgren states (p. 93) "The most fundamental changes in rocks take place with practical constancy of volume," he adds in a footnote, "When a rock is permeated by strong, rapidly moving liquids, solution may overtake deposition and a drusy structure may result. But even in this case the bulk volume is likely to remain about constant, except under very heavy pressure." This last statement says, in effect, that the rock volume has been *reduced*, leaving an opening equal in volume to that originally occupied by solid rock, and if the pressure of overlying ground is not too great this opening remains. Locke's idea apparently is that after this corrosive action by hydrothermal solutions, the pressure in many cases is too great and the rock overlying the cavity collapses into it, forming a breccia pipe, or else there is incipient collapse to form a shatter pipe. Such shatter pipes in closely packed swarms may well constitute the most common form of porphyry copper structure.

Another significant contribution on the relationship between breccia pipes and porphyry copper deposits was made by B. S. Butler (6). His discussion was an outgrowth of studies in the San Francisco District, Utah. Two of the pipes there, the Cactus and O. K. ore deposits, led Butler to review the literature on similar deposits. He cites other examples of such "quartz-tourmaline-chalcopyrite deposits" (p. 127–129), and concludes that both the Cactus and O. K. deposits "are . . . very closely allied to the type commonly known as disseminated deposits, of which there are numerous examples in the west" (p. 129). Such disseminated [porphyry copper] deposits: "occur in siliceous rocks, either schists, monzonites, or some closely allied rocks. These have been fractured and highly altered by ore-bearing solutions. The fissures are filled by veins of quartz and sulphides, and grains of sulphides are also disseminated through the altered rock. Both pyrite and chalcopyrite are usually abundant primary sulphides. The commercial ore is as a rule the result of enrichment" (p. 129).

Although there are no published field investigations to confirm the relationship between porphyry copper deposits and breccia pipes, it seems to the writer that what evidence is available is decidedly in favor of this concept. Besides the structural and mineralogic similarities pointed out by Locke and Butler, the following observations apply:

There are all gradations between deposits considered to be disseminated porphyry copper deposits and those considered to be breccia pipe deposits. For example, the Cactus "pipe" (Fig. 3) is hardly pipe-like in form and has shatter structure rather than breccia structure. It could be considered a small porphyry copper deposit or a shatter pipe that diverges greatly from a

pipe-like shape. The Childs-Aldwinkle deposit (Fig. 2) probably more closely approaches the typical co-hydrothermal breccia-shatter pipe than do most porphyry copper deposits, yet it is not essentially different from the Cactus deposit. The sulfide mineralization of the Childs-Aldwinkle deposit, including pyrite, chalcopyrite, and molybdenite as well as zones of pegmatite material, is certainly suggestive of porphyry copper mineralization.

The central, shattered part of the Pilares pipe resembles the fracturing at Bingham, and the Colorada pipe of Cananea, Mexico, is surrounded by a halo of porphyry copper-type mineralization. At Santa Rita, New Mexico, the relatively weakly mineralized ground near the center of the pit perhaps has a counterpart in the relatively weakly mineralized core of the Pilares deposit. The circular plan of the Climax, Colorado, molybdenum deposit with ore adjacent to a sheath-like mass of silica rock resembles features of the Colorada pipe deposit.

Other porphyry copper deposits, such as Bingham and Bagdad, are surrounded by relatively barren breccia pipes. If these breccia pipes, as well as the porphyry copper deposit, are regarded as having been formed by hydrothermal fluids, and if hydrothermal activity is considered to have been centered beneath the porphyry ore deposit, then the isolated breccia pipes logically represent the fringe effects of mineralizing activity as well as the fringes of a center of collapse caused by the corrosive action of the fluids. At this center, breccia or shatter columns coalesce, forming porphyry copper structure, and away from it, the columns, being less crowded together, become recognizable as separate entities.

The intense alteration around most porphyry copper deposits testifies to the activity and volume of hydrothermal fluids. The complex of structural intersections commonly centering on these deposits (36) perhaps provided the necessary concentration of channelways for the mineralizing fluids to effectively corrode large enough blocks of ground to cause the widespread shattering. Also, if the shattering in porphyry copper deposits is related through mineralization stoping to ore solutions characterized by a certain degree of corrosional activity, assuming that the rock is not abnormally reactive or non-reactive, as for example limestone and quartzite respectively, then perhaps this explains why shatter or stockwork structure is especially typical of copper-molybdenum deposits and not of, say, lead-zinc deposits. In other words solutions that deposit copper and molybdenum may be characterized by an ability to corrode the country rock to a degree (neither too great nor too small) that is most likely to induce porphyry copper-type shattering.

SUMMARY AND CONCLUSIONS

Breccia and pebble columns associated with epigenetic ore deposits present a wide variety of phenomena whether they are considered from the structural or mineralogical standpoint, or from the standpoint of origin. The broadest classification based on origin would include three main types distinguished on the basis of dominant movement of fragments. The movement of fragments generally is predominantly one of the following: 1. Intrusive;

2. Collapse; 3. Oscillatory. Intruded fragments generally are impelled by magma or else by gaseous or aqueous fluids that arise from the magma or are mobilized by magmatic heat. The fragments that constitute the column do not reach the surface and they accumulate either in a volcanic vent or in a hypabyssal intrusive body.

Collapse structures probably have an even wider variety of causes. Besides collapse caused by withdrawal of magma or slackening of intrusive activity, collapse might result also from melting of rock by gases compressed by an advancing magma column (52, p. 316), by distension of the crust caused by doming, by solution of limestone by groundwater, by shrinkage due to corrosion by gaseous or hydrothermal fluids, and possibly also by shrinkage due to cooling of an intrusive.

Oscillatory movement would result from a combination of intrusive and collapse processes.

These three types of movement, namely, intrusive, collapse, and oscillatory, produce rock masses in which the fragments are notably displaced, rotated and, in some cases, mixed. Incipient intrusion, collapse or oscillation on the other hand causes shatter structures in which the broken fragments are essentially in place relative to each other. Probably collapse processes are most important in producing shatter structures.

Breccia and pebble columns display a wide variety of shapes and internal structural features and a few dominant structural controls. They vary from dike-like to pipe-like and to forms that have no readily-described counterpart. Common internal features are zonal arrangements with respect to the pipe axis, of fragment size, fragment rounding, porosity, and fracturing.

Breccia and pebble columns are most common in competent rocks and rocks that do not have distinct directional properties. Within such rocks they generally show control by planar structures and structural intersections, and in many cases are apically situated with respect to instrusive bodies.

From the standpoint of mineralogy and economic geology, breccia and pebble columns fit into two broad categories termed *pre-hydrothermal* and *co-hydrothermal*. Pre-hydrothermal structures formed prior to the advent of hydrothermal fluids, and show relatively poor spatial coincidence with contained ore deposits. Co-hydrothermal structures formed as a result of corrosive action by hydrothermal fluids, show a relatively close coincidence with ore, and as a class probably display a higher incidence of ore mineralization than do the pre-hydrothermal class.

Most porphyry copper deposits probably are broad co-hydrothermal shatter columns, or aggregates of such columns, because structurally as well as mineralogically they show a close relationship to co-hydrothermal breccia pipes.

PROBLEMS RELATING TO BRECCIA AND PEBBLE COLUMNS

The subject of breccia and pebble columns contains enough debatable points to show that from many aspects it needs additional study. Data are needed to verify or to revise some of the concepts presented above. There is room for considerably more study to determine the exact nature of the

geologic processes involved in pipe formation, to find criteria for distinguishing between pre-hydrothermal and co-hydrothermal, as well as other types, and to find criteria for evaluating pipe structures as guides to ore. Another problem is that of terminology and classification, although recent progress is shown in Fisher's (16) proposed classification of volcanic breccias. Present usage indicates little or no grasp of the many possible ways in which column-like breccia structures can originate. The relationship of "mineralization stoping" to porphyry copper structure surely deserves more attenuation than it has received in the years that have elapsed since Locke suggested in 1926 that this process was responsible for porphyry copper shattering. More theoretical questions also offer an intriguing challenge. For example, are pipe-like structures, such as the one at Toquepala, really diatremes or are they, perhaps, more closely allied to hydrothermal processes than to volcanism? This leads to the broader question of the relationship between volcanism and hydrothermal ore deposition.

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COPPER, VANADIUM, AND URANIUM DEPOSITS IN SANDSTONE—THEIR DISTRIBUTION AND GEOCHEMICAL CYCLES [1]

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ABSTRACT

Deposits of copper, vanadium, and uranium in nonmarine sandstones are numerous and widespread. Copper deposits, with or without uranium, are mainly resident in first-generation arkosic sandstones derived from granitic rock terrains; deposits rich in vanadium, with or without much uranium, are dominantly in second-generation sandstones derived from sedimentary rocks; and the uranium deposits with little or no vanadium or copper are in either first- or second-generation sandstones, many of which are associated with beds containing volcanic debris.

All three metals are dispersed in igneous rocks but not in close association. Copper and uranium enter the hydrothermal environment, but the record of vanadium in hydrothermal solutions and veins is scant. Some of the uranium and most of the copper minerals in igneous rocks and veins oxidize readily and the metals go into surface- and ground-water solutions, but the vanadium in igneous rocks is not so easily mobilized—under normal geologic conditions, conceivably it may require diagenetic reactions and a second period of weathering to solubilize much vanadium. All three metals precipitate from solutions in the presence of a reducing agent, such as carbonaceous material or associated sulfide ions, either in sediments as they accumulate or in existing rocks.

These geochemical habits permit the concept that copper and uranium are made available by weathering of igneous rock terrains and hence might accumulate in first-generation sediments, whereas vanadium would be commonly available only after a second period of weathering. Perhaps the oxidation or devitrification of volcanic debris may contribute uranium to ground waters as does the weathering of igneous rocks.

INTRODUCTION

SANDSTONE deposits containing conspicuous accumulations of copper, vanadium, and uranium are numerous and widespread in western United States

[1] Publication authorized by the Director, U. S. Geological Survey.

and a few are in the eastern states. Some deposits contain only one of these metals, others contain both copper and uranium or vanadium and uranium. The uranium- and vanadium-bearing deposits have been the principal domestic source of these metals and they have been studied intensively in recent years; the copper deposits have not been very productive but are well known geologically as the red beds type.

Although these deposits have many common habits, the recent studies suggest a selective relationship between the metal content of the deposits and the lithologic and genetic characteristics of the host rocks. The copper deposits are mainly resident in first-generation arkosic sandstones derived from granitic rocks; deposits rich in vanadium, with or without much uranium, are dominantly in second-generation quartzose sandstones derived from sedimentary rocks; and the uranium deposits with little or no vanadium or copper are either in first- or second-generation sandstones, many of which are associated with beds containing altered volcanic debris.

This paper examines the distribution of these metals relative to their geochemical cycles.

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HABITS OF THE DEPOSITS

The ore minerals below the zone of oxidation are mainly sulfides of copper and low-valence oxides and silicates of vanadium and uranium—these metals do not combine with one another in these minerals. Accessory pyrite and marcasite are common but generally not abundant. In the zone of oxidation many secondary minerals form, some of which contain two of these metals. The ore minerals mainly fill pores in the sandstone; replacement of the host material is relatively minor except for rich and spotty concentrations in plant fossils and sandstone. Introduced gangue minerals other than those that commonly cement sandstone are inconspicuous or absent.

Typically the deposits are lenticular or tabular and nearly parallel to bedding. Many of them are small but some contain more than a million tons of ore. The influence of sedimentary structures, such as lenses, in localizing deposits has been recognized by many geologists; tectonic structures, on the other hand, have not obviously localized many deposits. Some deposits are near igneous intrusive bodies, and some are near hydrothermal veins, but these do not differ conspicuously from those that are remote from known centers of igneous or hydrothermal activity.

Although many geologists believe that these deposits represent a particular type having a similar origin, no single genetic hypothesis is generally accepted. Some geologists favor a hydrothermal introduction of the metals from a hypogene source. Others favor secretion from dispersed sources, either from the rocks from which the host sandstones were derived, laterally from the host sandstones, or vertically from adjacent stratigraphic units, the trans-

porting medium being conceived either as surface waters, connate waters, or hypogene solutions.

CHARACTERISTICS OF THE HOST SANDSTONES AND DISTRIBUTION OF METALS

The metal distribution and the geologic characteristics of the principal host rocks are described below; deposits in beds of Triassic age, where relations are somewhat complex, are discussed separately. Table 1 lists the stratigraphic and geographic distribution of nearly all sandstone deposits of

Table 1. Distribution and characteristics of the sandstones that are hosts to deposits of copper, vanadium, and uranium

Metal class	System	Group or formation and member	Dominant location, lithologic characteristic, and condition of accumulation	Genetic classification 1/		Approximate number of deposits 2/	Other remarks
				1st generation	2nd generation		
Copper and copper-uranium	Triassic	Moenave, Springdale	Utah; arkosic SS; stream	+ ?	- ?	few	Ag and V
		Dockum 3/	N. Mex.; SS, CGL; stream	- ?	+ ?	few	
		Chinle: Agua Zarca	N. Mex.; SS, CGL; stream	- ?	+ ?	10	
		Chinle: Shinarump 3/	Utah; CGL SS; stream	- ?	+ ?	50	
		Santa Rosa	N. Mex.; SS; stream	- ?	+ ?	few	
	Permian	Double Mountain (of former usage)	Okla., Texas; arkosic SS; stream	+		Many red beds copper deposits, some U	
		Clear Fork	Texas; arkosic SS; stream	+			
		Wichita	Okla., Texas; arkosic SS; stream	+			
		Abo	N. Mex.; arkosic SS; stream	+			
		Cutler	Utah; arkosic SS; stream	+			
	Permian and Pennsylvanian	Maroon	Colo.; arkosic SS; stream	+			
		Sangre de Cristo	N. Mex., Colo.; arkosic SS; stream	+			
	Devonian	Catskill	Pa.; graywacke; stream	+		40	
Vanadium-uranium	Cretaceous	Dakota 3/	N. Mex., Colo.; SS; stream		+	10	
		Inyan Kara 3/	S. Dak., Wyo.; SS; stream		+	100	
	Jurassic	Morrison: Brushy Basin 3/	Colo., Utah; SS, CGL; stream		+	20	Beds adjoining are somewhat tuffaceous
		Morrison: Salt Wash	Colo., Utah, Ariz.; SS; stream		+	2400	
		Entrada	Colo., N. Mex.; SS; wind		+ ?	few	Sparse U
		Kayenta	Ariz.; SS; stream		+	few	
		Navajo(?)	Colo.; SS; wind		+ ?	few	Sparse U
	Triassic	Chinle: Moss Back 3/	Utah; CGL SS; stream	- ?	+ ?	20	Beds adjoining are tuffaceous
		Chinle: Shinarump 3/	Ariz., Utah; CGL SS; stream	- ?	+ ?	10	
Uranium deposits with little or no vanadium or copper	Pliocene	Coso	Calif.; arkosic SS; stream	+		few	
		Bidahochi	Ariz.; tuffaceous SS; lake		+ ?	few	
		Goliad	Texas; SS; stream, shoreline		+ ?	1	
	Miocene	Browns Park	Colo., Wyo.; arkosic SS; stream	+		30	
	Oligocene	White River	Wyo.; tuffaceous SS; stream	+		20	
	Eocene	Jackson	Texas; tuffaceous SS, stream, shoreline		+ ?	10	
		Tepee Trail	Wyo.; arkosic SS; stream	+		10	Beds associated are commonly tuffaceous
		Wasatch	Wyo.; arkosic SS; stream	+ ?	- ?	400	
		Wind River	Wyo.; arkosic SS; stream	+		100	
	Paleocene	Fort Union	Wyo.; SS; stream, lake	- ?	+ ?	20	
	Tertiary undivided		Nev., Colo.; arkosic SS; stream, lake	+ ?	- ?	100	
	Cretaceous	Mesaverde: Point Lookout	N. Mex.; SS; stream, shoreline	- ?	+ ?	10	
		Mesaverde: Toreva	Ariz.; arkosic SS; stream, shoreline	+		10	Sparse V
		Dakota 3/	N. Mex., Colo.; SS; stream		+	10	Sparse V
		Inyan Kara 3/	S. Dak., Wyo.; SS; stream		+	100	Sparse V
	Jurassic	Morrison: Brushy Basin 3/	N. Mex.; arkosic SS; stream	+ ?	- ?	10	Beds adjoining are tuffaceous
		Morrison: Westwater Canyon	N. Mex.; arkosic SS; stream	+ ?	- ?	30	
		Morrison: Recapture	Utah; arkosic SS; stream	+ ?	- ?	few	
	Triassic	Newark	Pa., Conn.; arkosic SS; stream	+		few	
		Dockum 3/	N. Mex.; SS, CGL; stream	- ?	+ ?	20	Sparse Cu
		Chinle: Petrified Forest 3/	Ariz.; tuffaceous SS; stream	- ?	+ ?	20	Beds adjoining are tuffaceous
		Chinle: Moss Back 3/	Utah; CGL SS; stream	- ?	+ ?	60	
		Chinle: Moniter Butte	Utah; SS; stream	- ?	+ ?	10	
		Chinle: Shinarump 3/	Ariz., Utah; CGL SS; stream	- ?	+ ?	80	
	Pennsylvanian	Tensleep	Wyo., Mont.; SS; shoreline		+ ?	few	Sparse V
		Pottsville	Pa.; arkosic SS, CGL; stream	+		1	
	Mississippian	Mauch Chunk	W. Va.; SS; shoreline	+ ?		1	

1/ Plus (+) sign indicates wholly or dominantly of the genetic classification shown; minus (-) sign indicates subordinate part; question mark(?) indicates uncertainty.

2/ The numerical distribution of deposits recorded by W. I. Finch is expressed in this table by quantitative words or rounded numbers to indicate only orders of magnitude.

3/ Stratigraphic unit also present in another metal class.

these metals and summarizes the characteristics of the host beds. The genetic classification of these beds is largely the responsibility of the writers; the rest of the tabulated information was taken from material compiled from many published and unpublished sources by Warren I. Finch for another report, a preliminary version of which has been published (7).

The host sandstones are chiefly nonmarine; most accumulated by stream action but a few were formed by wind or under lake or shoreline conditions. Some deposits are in beds composed of sands that were not transported far from their source areas; others are at sites 100 miles or more from the sources of sand. The host sandstones are typically lenticular and commonly interbedded with mudstone. Carbonized plant fossils are conspicuous in many of the sandstones, and asphaltic and petroliferous material is present in some, but a few of the host beds contain no recognized carbonaceous material. Altered volcanic debris is abundant in mudstones and sandstones associated with some of the ore-bearing beds, and a few of the host sandstones are tuffaceous.

The range in composition of the principal ore-bearing sandstones is shown on triangular diagrams (Fig. 1) by grouping the detrital components as they are diagnostically associated in tuffs, arkoses, and quartzose sandstones. One group consists of tuffaceous material and montmorillonite, representing the contribution of volcanic debris; the second consists of feldspar, kaolinite, illite, and mica, which are commonly derived by the erosion of igneous and metamorphic rocks; and the third consists of quartzite and chert, which are commonly derived from pre-existing sediments, plus quartz, which of course may be derived from sedimentary or igneous and metamorphic rocks.

The few available samples from host rocks containing copper and copper-uranium deposits are in a narrow field along the arkose-quartzose sandstone line (Fig. 1-A) and would be classed as arkose and feldspathic orthoquartzite by Pettijohn (23). Samples from formations containing vanadium-uranium deposits are clustered toward the quartzose sandstone corner (Fig. 1-B) and would be classified as orthoquartzite and feldspathic orthoquartzite. Samples from formations containing deposits of uranium with little or no vanadium or copper (Fig. 1-C) have a wider spread from the quartzose sandstone corner toward arkose and tuff and would be classified as orthoquartzite, feldspathic orthoquartzite, arkose and tuffaceous sandstone.

Most of the copper deposits are in coarse-grained arkosic sandstones, some of which are conglomeratic; a few are in shaly beds. The mean grain size of the samples studied is 0.33 mm, but the samples are too few to consider this figure significant. Practically all of the vanadium-rich deposits are in sandstone of fine- to medium-grain size, and the samples show a mean of 0.14 mm. Samples of formations containing the deposits of uranium with little or no vanadium or copper also have a mean grain size of 0.14 mm, but the host rocks of these deposits range from silty beds to conglomerates.

The grain-size distribution of many of the samples used in preparing Figure 1 has been analyzed to determine the degree of sorting of the sandstones represented. The logarithmic standard deviation expressed in phi units ranges 0.9 to 1.7 and averages about 1.2 for the sandstones that contain

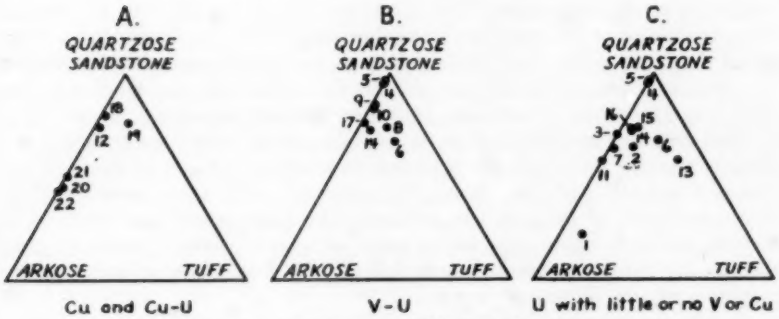

FIG. 1. Diagrams showing composition of the host sandstones.

Sample 1, triangle C. Browns Park formation, 2 samples (R. A. Cadigan, communication).

2, C. Wasatch formation, estimate from thin sections (W. N. Sharp, communication).

3, C. Wind River formation, estimate from thin sections (R. G. Coleman, communication).

4, B and C. Inyan Kara group, 115 samples (1).

5, B and C. Dakota sandstone, 6 samples (R. A. Cadigan, communication).

6, B and C. Brushy Basin member, Morrison formation, 15 samples (R. A. Cadigan, communication).

7, C. Westwater Canyon member, Morrison formation, 23 samples (R. A. Cadigan, communication).

8, B. Salt Wash member, Morrison formation, 23 samples (R. A. Cadigan, communication).

9, B. Entrada sandstone, 18 samples (R. A. Cadigan, communication).

10, B. Navajo sandstone, 9 samples (3).

11, C. Newark group (21).

12, A. Springdale sandstone member, Moenave formation, 3 samples (2).

13, C. Petrified Forest member, Chinle formation, 2 samples (3).

Sample 14, B and C. Moss Back member, Chinle formation, 6 samples (3).

15, C. Monitor Butte member, Chinle formation, 17 samples (3).

16, C. Shinarump member, Chinle formation, Circle Cliffs area, Utah, 5 samples (3).

17, B. Shinarump member, Chinle formation, Monument Valley area, Ariz., 10 samples (3).

18, A. Shinarump member, Chinle formation, White Canyon area, Utah, 7 samples (3).

19, A. Agua Zarca sandstone member, Chinle formation, 10 samples (R. A. Cadigan, communication).

20, A. Cutler formation, 4 samples (R. A. Cadigan, communication).

21, A. Abo formation, 2 samples (J. H. Stewart).

22, A. Sangre de Cristo formation (25).

vanadium-uranium deposits. The standstones that contain deposits of uranium with little or no vanadium or copper range from 1.3 to 2.6 and average about 1.85 phi units, and those that contain copper and copper-uranium deposits range from 1.7 to 2.1 and average 1.9, indicating much poorer sorting than in the sandstones containing vanadium-uranium deposits.

Red beds copper deposits (6, 9, 14, 25), some containing uranium and some a trace of vanadium, are abundant in some sandstones of Permian and Triassic ages in Oklahoma, Texas, New Mexico, and Utah (Table 1). The host sandstones of Permian age are especially arkosic and were derived, at least to a large extent, from source areas of igneous rocks. Those hosts of Triassic age are less conspicuously arkosic and second-generation material may be a major component in some.

Vanadium-uranium deposits are abundant in the Salt Wash member of the Morrison formation of Jurassic age in western Colorado and eastern Utah. Similar deposits occur in the Inyan Kara group of Cretaceous age in the Black Hills area of South Dakota and Wyoming, though none are as rich in vanadium as those in Colorado and Utah, and some contain very little vanadium. The sandstones in the Salt Wash are fine- to medium-grained, sparsely arkosic and tuffaceous, and were derived mainly from a terrain of sedimentary rocks in west-central Arizona and southeastern California (4); sandstones in the Inyan Kara are fine-grained, rich in quartz, and were derived mainly from sedimentary rocks to the southeast of the Black Hills (1). Beds containing some altered volcanic ash are associated with the ore-bearing sandstones in both areas.

A few large vanadium deposits, with only a little uranium, are in the Entrada and Navajo(?) sandstones of Jurassic age in western Colorado. These sandstones are well sorted, they contain much quartz but little arkosic material, and probably were derived from older sandstones. Immediately adjacent beds contain little or no volcanic debris.

Large uranium deposits that contain little or no vanadium or copper occur in the Westwater Canyon and Brushy Basin members of the Morrison formation in New Mexico (18), in several formations of Tertiary age in Wyoming (17), and in the Jackson group (5) and Goliad sand, both of Tertiary age, in Texas. The deposits in New Mexico are near the southern edge of Morrison accumulation, and they are in arkosic sandstones presumably derived, at least in part, from igneous rocks to the south. Associated with the ore-bearing sandstones are bentonitic mudstones rich in montmorillonite. The ore contains a little vanadium. The ore-bearing Tertiary sandstones in Wyoming accumulated in intermontane basins bordered by upland masses in which igneous rocks were exposed. The Wind River formation, which is ore-bearing in the Gas Hills and the Shirley Basin, and the equivalent Wasatch formation, which is ore-bearing at Crooks Gap and in the Powder River Basin, are arkosic. These units are overlain by tuffaceous shales and sandstones, and the Wasatch itself contains some tuffaceous material. The Browns Park formation, which has yielded ore from deposits at Baggs, Wyo., and near Maybell, Colo., is arkosic and contains tuffaceous material. The

host beds in Texas are composed of sands and muds derived from sedimentary rocks, and they are tuffaceous and bentonitic.

Several stratigraphic units of Triassic age are ore-bearing in Utah, Arizona, and New Mexico. Relations between the metal distribution and the characteristics of the host beds in these units are not as consistent as those described above. The metal composition of the deposits differs more from place to place than in deposits in rocks of other geologic ages. The Triassic host beds, on the other hand, are rather similar in composition (Fig. 1, samples 12 and 14–18) even though these units are widespread and probably are composed of material derived from different source areas. The metal distribution may be related in part to these different sources.

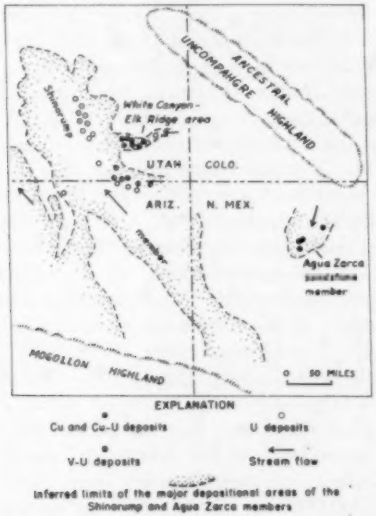

FIG. 2. Map showing general distribution and type of ore deposits in the Shinarump member and Agua Zarca sandstone member of the Chinle formation.

In northeastern Arizona and south-central Utah the Shinarump member of the Chinle formation contains uranium and vanadium-uranium deposits (Fig. 2), some of which contain a little copper. The Shinarump in this area is a broad, relatively thin layer of conglomeratic sandstone that was derived mainly from rocks predominantly volcanic and sedimentary in the Mogollon Highland to the south. In contrast, copper-uranium and uranium deposits occur in a rather well-defined arm of the Shinarump in the White Canyon–Elk Ridge area, Utah (Fig. 2). The host rock in this area was deposited by westerly flowing streams (19), and although this part of the Shinarump is no more arkosic than other parts (Fig. 1), it probably received some material from the granitic and metamorphic rocks of the Ancestral Uncompahgre

Highland. In north-central New Mexico the Agua Zarca sandstone member of the Chinle is probably equivalent to the Shinarump, and it contains copper deposits, some with a little uranium. The Agua Zarca was laid down by southwesterly flowing streams (Fig. 2), and probably some of its composing material was derived from Precambrian rocks (11) at the southeast end of the Ancestral Uncompahgre Highland. Thus the abundance of copper in deposits in the Shinarump of the White Canyon–Elk Ridge area and in the Agua Zarca may be related to the igneous and metamorphic rocks that contributed material to these units, whereas the uranium and vanadium-uranium deposits in the Shinarump of Arizona and south-central Utah may be related to the volcanic and sedimentary rocks that contributed material to this part of the Shinarump.

The Dockum group in eastern New Mexico is generally correlated with the Chinle, and it also contains copper- and uranium-bearing deposits. Some of its composing material may have been derived from remnants of the southern end of the Ancestral Rocky Mountain, which during Permian time was the source of arkoses that contain copper deposits of the red-beds type.

Other members of the Chinle formation also contain ore deposits in Utah and Arizona (8). The Monitor Butte member is host to uranium deposits in central Utah, the Moss Back member to vanadium-uranium and uranium deposits in central and southeastern Utah, and the Petrified Forest member to uranium deposits in north-central Arizona. Some of these deposits contain a little copper. All of these sandstones are about as arkosic as is the Shinarump; in addition, the Petrified Forest member is tuffaceous, and all of the ore-bearing sandstones in the Chinle are associated with beds that contain altered volcanic debris. The Monitor Butte, Moss Back, and Petrified Forest members of the Chinle were derived chiefly from volcanic and sedimentary rocks in the Mogollon Highland to the south (3, p. 523).

In southwestern Utah the Springdale sandstone member of the Moenave formation, which is younger than the Chinle, contains deposits classed as copper-uranium in Table 1, but these deposits have yielded a greater production value (about $8,000,000) in silver. A little vanadium is also present. The Springdale member is about as arkosic as the Shinarump (Fig. 1-A).

GEOCHEMICAL CYCLES OF THE METALS

Information in this section is summarized largely from Rankama and Sahama (24); Goldschmidt (13); Krauskopf (20); McKelvey, Everhart, and Garrels (22); and Gruner (16). Estimates of the average content of copper, vanadium, and uranium in common rock types are shown in Table 2.

Copper averages about 70 ppm in igneous rocks. It occurs mostly as sulfides (mainly chalcopyrite) dispersed among rock-mineral grains, but some enters silicate minerals and iron oxides, perhaps in substitution for iron. Although most of the copper in magmas concentrates in mafic rocks, some precipitates in silicic rocks, and some is available for hydrothermal transport. Copper sulfides in igneous rocks and veins oxidize readily and the copper is mobilized. Because of an affinity for clay minerals, copper, in small concen-

trations, can be carried by mud suspended in surface waters, but some copper enters the surface- and ground-water systems in solution. Glagoleva (12) shows an average of about 100 ppm copper in the suspended material in eight samples of water from Black Sea basin rivers; in all but one sample, however, the total amount of copper carried in solution was greater than that in the associated suspended material. The average copper content of shales is in the same order as that of igneous rocks whereas the average content of sandstones and limestones is somewhat smaller (Table 2). Copper can be moderately to richly concentrated from surface- or ground-water solutions in the reducing environments of iron sulfides or carbonaceous material, either in existing rocks or in sediments that are accumulating.

TABLE 2

AVERAGE CONTENTS OF COPPER, VANADIUM, AND URANIUM IN COMMON ROCK TYPES

[In parts per million]

Metal	Igneous rocks					Sedimentary rocks			Reference
	Ultramafic	Mafic	Intermediate	Silicic	Average of igneous rocks	Shales	Sandstones	Limestones	
Copper	80	140	35	30		57			Vinogradov (26)
	80	145	35	20	70	38–90	10–40	6–18	Green (15)
		149	38	16	70	192		20	Rankama and Sahama (24)
					70	30–150	10–40	5–20	Krauskopf (20)
Vanadium	140	200	100	40		130			Vinogradov (26)
	140	210	95	55	90	90–130	10–60	2–20?	Green (15)
					150	50–300	10–60	2–20?	Krauskopf (20)
Uranium	0.03	0.8	1.8	3.5		3.2			Vinogradov (26)
	0.3	0.8	2.1	3.8	2.6	3.7–4.1	0.45	1.9–2.5	Green (15)
		0.9	2.6	4	4	1.2	1.2	1.3	Rankama and Sahama (24)

Vanadium averages about 100 ppm in igneous rocks; it is more abundant in mafic rocks than silicic ones. It does not form the essential part of any rock mineral but rather its ions replace those of iron and aluminum in dark-colored silicates and iron and titanium in oxides; vanadium is especially enriched in magnetite in titaniferous magnetite deposits. Most of the vanadium in magmas is apparently fixed in minerals in the relatively insoluble trivalent state at an intermediate stage of magmatic differentiation, and not much is available for hydrothermal transport (10). During the normal weathering of igneous rocks, the vanadium that is released (mostly from silicate minerals) is said to go into the clay minerals formed, and it probably remains with these minerals as they are transported to places where they accumulate as sedimentary rocks, for argillaceous sediments are generally more vanadiferous than other types—vanadium averages about 100 ppm in shales (about the same content as igneous rocks), perhaps about 30 ppm in sandstones, and

10 ppm in limestones. Glagoleva (12) shows about 100–150 ppm vanadium in the suspended material of the waters from Black Sea basin rivers, but no vanadium was detected in solution in any sample.

If, on the other hand, igneous rocks or vanadiferous clay minerals are weathered under conditions of intensive oxidation, as in an arid climate, the low-valence vanadium is converted to the soluble quinquevalent state and goes readily into surface- and ground-water solutions. From these solutions vanadium can be precipitated and locally concentrated, either in existing rocks or in sediments as they accumulate, in the presence of hydroxides of aluminum or ferric iron; by cations of heavy metals such as lead, zinc, and copper; or in a reducing environment, as in the presence of carbonaceous material.

Uranium averages about 3 ppm in igneous rocks, and it concentrates more abundantly in the silicic part of igneous differentiates. Some of it is held in accessory, refractory minerals that do not oxidize readily, and some is in a loosely bound form in mineral interstices. Uranium, mainly as uraninite, is common in pegmatites and hydrothermal veins. On weathering, uraninite oxidizes readily, forming soluble salts and secondary minerals, from which uranium can enter surface and ground waters. The loosely bound uranium in igneous rocks also can be introduced into these waters. Uranium averages about 3 ppm in shales, ½ ppm in sandstones, and 2 ppm in limestones, but if organic material or sulfide ions are present in these rocks and create a reducing environment, more uranium may concentrate. This process of reduction and concentration evidently can occur as sediments accumulate, for example in black shales and perhaps coals, or perhaps it can take place from solutions percolating through rocks containing organic remains and pyrite, such as in some sandstones. The concentrations and habits of uranium in volcanic rocks and ash are not well known, but certainly some uranium is present in some volcanic material, and presumably it could be made soluble by oxidation during and after accumulation of volcanic debris in sedimentary rocks and perhaps during devitrification of ash. If uranium in the volcanic material is mobilized, it could enter surface or ground waters and be carried to other rocks.

SUMMARY AND INTERPRETATIONS

Copper deposits of the red-beds type are dominantly in first-generation sandstones, vanadium deposits are mainly in second-generation sandstones, and uranium deposits occur in sandstones of both first- and second-generation origin. Volcanic debris is common in beds associated with the host sandstones of many uranium deposits, especially those that are of second-generation origin. The geochemical cycles of these metals permit an explanation of this distribution.

All three metals are dispersed in igneous rocks, but not in close association. Copper and uranium commonly enter the hydrothermal environment, but the record of vanadium in hydrothermal veins is scant. The principal mineral forms of copper and some mineral forms of uranium in igneous rocks and veins oxidize readily to a higher valence state on weathering and the

metals are mobilized for transport by surface and ground waters and thus are available to circulate in first-generation sediments. Possibly volcanic debris in beds associated with host sandstones of second-generation origin might play the same role in contributing uranium as an igneous rock terrain might play in contributing to a first-generation sandstone. Vanadium, on the other hand, commonly remains in a relatively insoluble state, either with iron and titanium oxide minerals or with clay minerals, on the normal weathering of igneous rocks. Although it can be released by intensive oxidation of igneous rocks, perhaps it is commonly freed from normal weathering products of igneous rocks only in the more reactive diagenetic environments and under a second period of erosion, thus becoming available for migration in second-generation sediments. All three metals can be precipitated from their mobile, high-valence states, by a reducing environment, perhaps like that in the vicinity of carbonaceous material, which is commonly present in the host sandstones. Copper, and perhaps uranium also, can replace pyrite, which might have previously replaced fossil wood in these deposits, but the replacement of pyrite by vanadium minerals is not commonly reported.

Even though the geochemical cycles of these metals may permit an explanation of the metal distribution, this paper does not postulate an origin for these deposits, for the scope of this paper does not allow consideration of many other factors of genetic significance. Continued study of these relations, however, may ultimately contribute to a better understanding of genesis.

U. S. GEOLOGICAL SURVEY,
DENVER, COLORADO, AND MENLO PARK, CALIFORNIA,
Aug. 3, 1960

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ANION METASOMATIC REPLACEMENT REACTIONS[1]

L. L. AMES, JR.

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ABSTRACT

Data on the kinetic anion replacement systems calcite–PO_4^{-3}, calcite–F^-, gypsum–CO_3^{-2}, and gypsum–PO_4^{-3} are given, along with concomitant Sr^{+2} inclusion into the final product. In a system containing several anions, only the anion forming the least soluble compound with the available cation is stable. The replacement reaction rate is directly proportional to the solubility difference between the original solid and final product. Several other variables also affect the reaction rate including temperature, column flow rate, influent pH, surface area of the initial solid, concentration of extraneous ions, and concentration of active anion. These variables affect diffusion rates to and from reaction sites as well as differences in solubility between initial and final reaction products. The crystal structure of the resulting product greatly affects the concomitant removal rate of Sr^{+2} into the resulting product.

INTRODUCTION

SINCE the study of the kinetic anion replacement systems calcite–PO_4^{-3} (1, 3) and calcite–F^- (2), other similar systems were investigated. These include the geochemically interesting gypsum–CO_3^{-2} and gypsum–PO_4^{-3} systems.

These reactions are referred to as "anion" because only the anions are active during the replacement process. The cation of the original solid is retained in the replacement product along with other cations from the solution that can be accepted by the replacement product crystal lattice. The anion from the solution moves into the original solid, forming a product that represents a more stable compound in that environment. The anion of the less stable, initial compound moves into the solution. Rearrangement of the original crystal lattice occurs at an interface with the replacement product.

It is the purpose of this paper to investigate some of the general principles governing anion replacement reactions, and their relationship to concomitant cation removal into the same final product.

[1] Work performed under Contract No. AT (45-1)-1350 for the U. S. Atomic Energy Commission.

METHODS OF INVESTIGATION

The experimental apparatus is shown in Figure 1. Influent solutions were passed through Pyrex columns using a Lapp "Microflo" pump calibrated to deliver standardized flow rates through the columns. Temperatures were maintained with blade heaters controlled with a thermoswitch. Influent and effluent solutions were collected and submitted to the Analytical Laboratory of the Hanford Laboratories Operation for radioanalysis. The high-purity Sr^{85}, P^{32}, and C^{14} used in this study were obtained from Oak Ridge, Tennessee. The other chemicals used were of reagent grade in distilled water solutions.

To allow direct comparisons between cation and anion removals, a dimensionless index, C/C_0, was utilized. C_0 is the concentration of radioisotope in the influent solution and C is the effluent concentration.

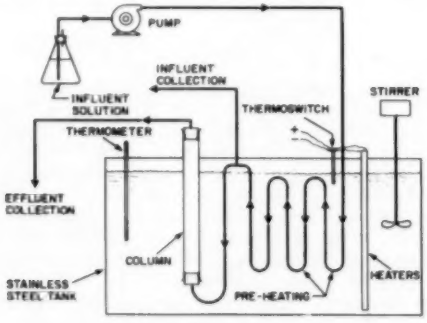

FIG. 1.

The experimental approach was designed to show the effect of independent variables (influent pH, flow rate, surface area of the initial solid, temperature, and concentration of the active anion) on the dependent variable (C/C_0). "Standard conditions" were arbitrarily chosen to assure a reasonable reaction rate. While one of the above-mentioned variables was held at standard conditions, the others were varied over ranges on both sides of these standard conditions. This procedure yielded a series of curves that theoretically should have passed through a single point at "standard conditions." The magnitude of this deviation from a single point represents total experimental error. Up to ± 10 per cent deviation can be expected at "standard conditions" because of the compounding of several weighing and measuring errors in obtaining each curve that tend to displace this single point horizontally rather than vertically.

RESULTS

Figure 2 shows a photomicrograph of an apatite layer formed on an optical grade calcite rhomb by immersion of the calcite in a 0.1M Na_3PO_4 solu-

FIG. 2.

tion for two weeks. Similar photomicrographs of the replacement of calcite by fluorite were presented elsewhere (2).

Figures 3 and 4 present the data on calcite–PO_4^{-3}–P^{32} removal and calcite–PO_4^{-3}–Sr^{85} removal. This reaction was discussed in detail elsewhere

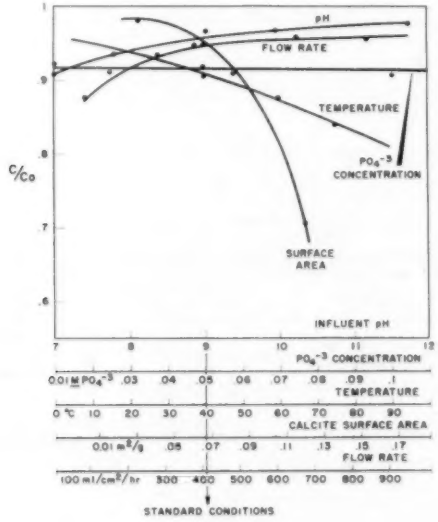

FIG. 3.

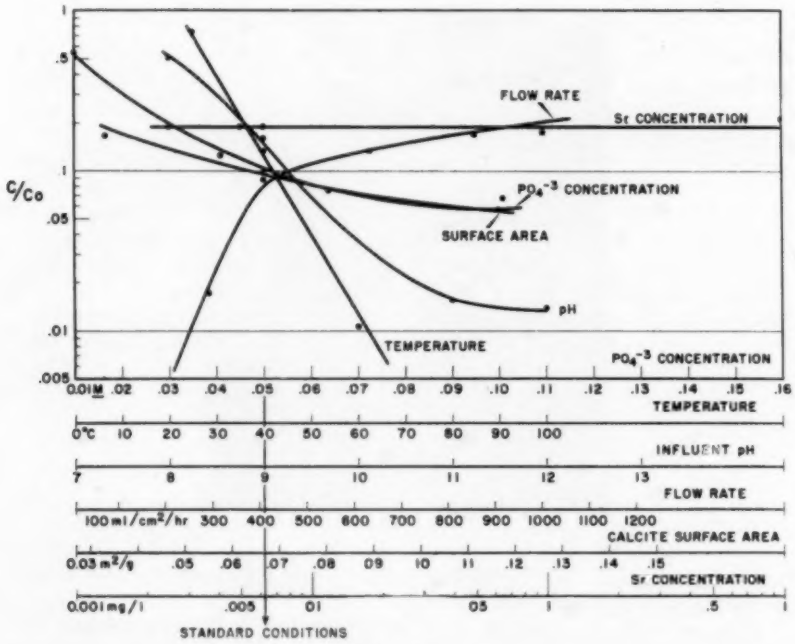
C/Co
FLOW RATE
Sr CONCENTRATION
PO4-3 CONCENTRATION
SURFACE AREA
pH
TEMPERATURE
PO4-3 CONCENTRATION
TEMPERATURE
INFLUENT pH
FLOW RATE
CALCITE SURFACE AREA
Sr CONCENTRATION
STANDARD CONDITIONS

FIG. 4.

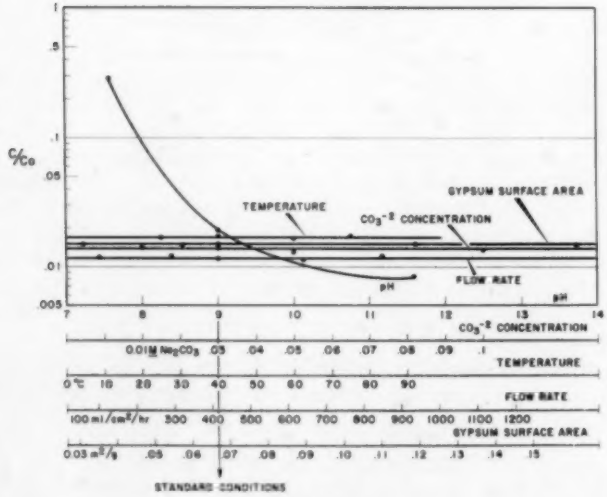
C/Co
GYPSUM SURFACE AREA
TEMPERATURE
CO3-2 CONCENTRATION
pH
FLOW RATE
CO3-2 CONCENTRATION
TEMPERATURE
FLOW RATE
GYPSUM SURFACE AREA
STANDARD CONDITIONS

FIG. 5.

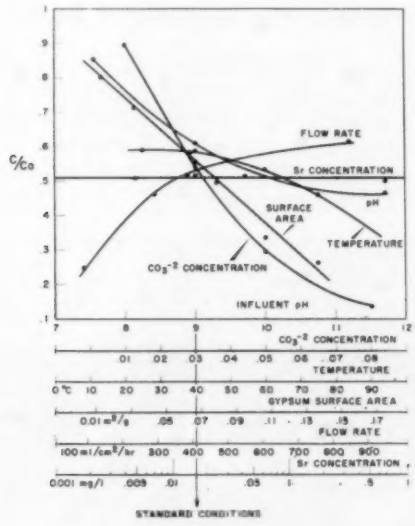

FIG. 6.

(1, 3), as were the calcite–F^- and calcite–F^-–Sr^{85} data utilized here (2). Note that the removal of Sr^{85} with pH in Figure 4 does not follow that for P^{32} with pH in Figure 3. Therefore, the author's previous contention that Sr^{90} could be used to trace the relative rate of apatite formation (1, p. 836)

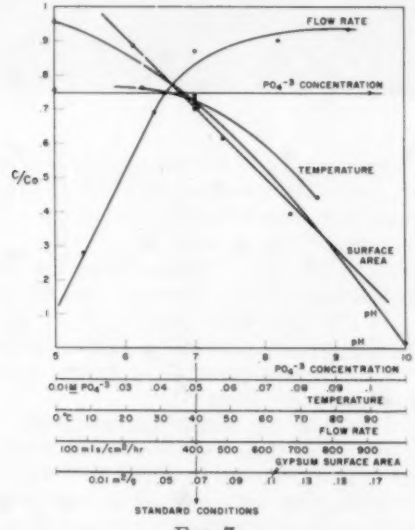

FIG. 7.

was erroneous. All of these solutions were 1 *M* NaCl in addition to their active constituents.

Figures 5 and 6 give data on the replacement of gypsum by calcite as traced with C^{14}, and concomitant Sr removal traced with Sr^{85}. Figure 7, the reaction of gypsum with PO_4^{-3}–P^{32}, is included to allow comparison of these P^{32} removal data with those of calcite in Figure 3. Note, though, that standard conditions include a pH of 7 with gypsum but a pH of 9 with calcite. The removal rate of PO_4^{-3} is depressed at pH 7 in comparison to pH 9.

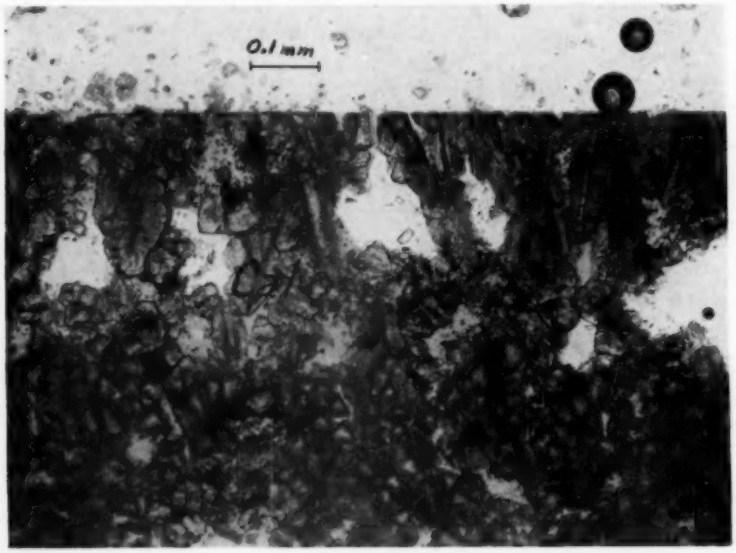

FIG. 8.

Figure 8 is a photomicrograph of a former single crystal of gypsum completely replaced with calcite by immersion in a 0.2 *M* Na_2CO_3 solution for two weeks. The apparent volume reduction may be due to material removed during cutting and grinding of this section.

A chemical analysis of the CaF_2 resulting from the alteration of calcite by alkaline, fluoride-bearing solutions showed less than 0.22 percent by weight CO_2. Compare this CO_2 value with up to 10 per cent by weight reported for apatite (1) formed under comparable conditions by alkaline PO_4^{-3}-bearing solutions. As was pointed out several times by McConnell (7), CO_3^{-2} does not substitute for F^- or OH^-.

Several runs were made in the kinetic system calcite–Na_2CO_3–Na_2SO_4–NaF–Na_3PO_4. The results of these experiments are presented in Table I.

The final products in all cases were identified by X-ray diffraction. Note that one obtains a final product, not a mixture of final products. The same experimental apparatus as shown in Figure 1 was used to obtain these results.

DISCUSSION

The reason that metasomatic replacement reactions proceed at all is that the original material being replaced is more soluble in the system than the final product. In anion replacement, the anion forms the least soluble compound

TABLE I

KINETIC PHASE RELATIONS IN SOME CALCIUM–ANION SYSTEMS AS AFFECTED BY pH

Influent solution	Influent pH	Effluent pH	Column material	Final product
0.03 *M* Na_2SO_4	3.1	6.6	Calcite	Calcite
0.03 *M* Na_2CO_3	10.9	10.8	Fluorite	Calcite
0.03 *M* Na_2SO_4	4.2	4.1	Fluorite	Fluorite
0.03 *M* Na_2SO_4	4.2	4.2	Apatite	Gypsum
0.03 *M* Na_2CO_3 0.03 *M* Na_2SO_4	4.1	6.5	Calcite	Calcite
0.03 *M* Na_2SO_4 0.03 *M* Na_2CO_3	9.2	9.2	Calcite	Calcite
0.03 *M* NaF 0.03 *M* Na_2CO_3	7.2	8.6	Calcite	Fluorite
0.03 *M* NaF 0.03 *M* Na_2CO_3	9.6	9.8	Calcite	Fluorite
0.03 *M* NaF 0.03 *M* Na_2CO_3	7.1	7.2	Gypsum	Fluorite
0.03 *M* NaF 0.03 *M* Na_2SO_4 0.03 *M* Na_2CO_3	9.3	9.7	Calcite	Fluorite
0.03 *M* NaF 0.03 *M* Na_2SO_4 0.03 *M* Na_2CO_3	9.2	9.2	Fluorite	Fluorite
0.03 *M* NaF 0.03 *M* Na_2SO_4 0.03 *M* Na_2CO_3 0.03 *M* Na_2PO_4	11.8	11.7	Calcite	Apatite
0.03 *M* NaF 0.03 *M* Na_2SO_4 0.03 *M* Na_2CO_3 0.03 *M* Na_2PO_4	7.2	7.4	Calcite	Apatite
0.03 *M* NaF 0.03 *M* Na_2SO_4 0.03 *M* Na_2CO_3 0.03 *M* Na_2PO_4	4.2	6.5	Calcite	Fluorite

The data of Table I were obtained by passing the influent solutions through two centimeter diameter chromatographic tubes containing three grams of reagent grade initial column material at 60° C.

with the cation or cations of the available solid. If the initial solid is the least soluble in a given system, there will be no replacement. No solid phases will result when all possible phases are highly soluble in aqueous solutions, or, as is the case with $MgCO_3$–F^-, the least soluble compound (MgF_2) does not form at room temperature. For example, if water is

passed through a column containing NaCl, no replacement reaction occurs because all possible replacement products in the system (NaCl, NaOH, HCl) are highly water-soluble. Just as MgF_2 cannot be precipitated from alkaline, $NaF–MgCl_2$ solutions at room temperature, $MgCO_3$ cannot be replaced with fluoride to form MgF_2 under the same conditions.

The solubility difference in a given system between the original solid and a possible alteration product determines whether or not a given replacement reaction will occur, and greatly affects the rate at which the reaction will occur. Given several possible alteration products less soluble than the original solid, and assuming a completed reaction, only the least soluble

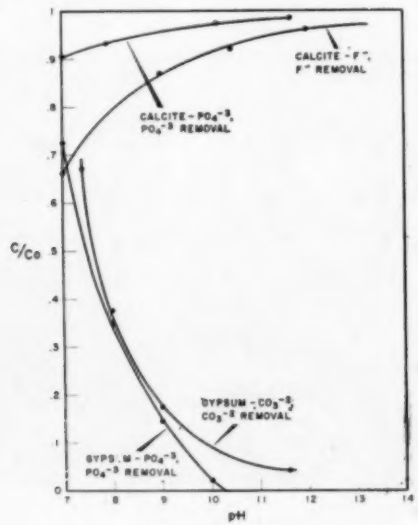

FIG. 9.

product is stable in the system. The data of Table I clearly show only one final product for each multiple anion system containing one cation. These data also suggest that the replacement of gypsum by calcite is an irreversible reaction, at least above pH 3.

Several variables will affect the difference in solubility between original and final product including active anion concentration, extraneous ion concentrations, temperature and pH. With systems containing hydroxyl and carbonate compounds, we may assume that solution pH would have a significant effect on this solubility difference. Figure 9 is a comparison of several anion replacement reactions under comparable conditions. Note that the specific reaction rates for calcite–PO_4^{-3} and calcite–F^- are decreasing respectively with increasing pH. This divergence with increasing pH is interpreted as a decreasing difference between the solubility of $CaCO_3$ and CaF_2,

and between the solubility of $CaCO_3$ and apatite. The difference in solubility between gypsum and calcite or apatite, on the other hand, increase with increasing pH, as do the gypsum–CO_3^{-2} and gypsum–PO_4^{-3} reaction rates.

These solubility differences can be approximated by the relative amounts of common cation (Ca^{+2}) that each (initial solid and final product) releases into solution under given equilibrium conditions. With solutions 0.03 *M* NaF, 0.03 *M* Na_2SO_4, 0.03 *M* Na_2CO_3 and 0.03 *M* Na_3PO_4, we can infer which solid is stable at several pH values. The computation for CaF_2 is straightforward. Assuming the HF is completely ionized we may use the activity product constant for $CaF_2 = 1.7 \times 10^{-10}$ (5). Given that $[F^-] =$ 0.03 *M*, the concentration Ca^{+2} at 25° C is 1.89×10^{-7} *M*. The computation for gypsum is likewise unaffected by $[H^+]$ above pH = 3, and with a gypsum activity product constant = 4.57×10^{-5} (5), yields a $[Ca^{+2}]$ of 1.52×10^{-3} *M* in 0.03 *M* Na_2SO_4 at 25° C. With calcite, $[CO_3]_T$, or total carbonate, does not equal $[CO_3^{-2}]$. Therefore, we must use the relationship

$$[CO_3^{-2}] = \frac{[CO_3]_T}{\frac{[H^+]^2}{K_{12}K_1} + \frac{[H^+]}{K_1} + 1},$$

where $[CO_3]_T$ = total carbonate = 0.03 *M*, K_1 = the dissociation constant for

$$HCO_3^- \ (7) = \frac{[CO_3^{-2}][H^+]}{[HCO_3^-]} = 6.02 \times 10^{-11},$$

and K_{12} = the dissociation constant for

$$H_2CO_3 \ (7) = \frac{[H^+][HCO_3^-]}{[H_2CO_3]} = 4.46 \times 10^{-7}.$$

Thus at pH 7 and 25° C,

$$[CO_3^{-2}] = \frac{3 \times 10^{-2}}{\frac{(1 \times 10^{-7})^2}{(6.02 \times 10^{-11})(4.46 \times 10^{-7})} + \frac{(1 \times 10^{-7})}{(6.02 \times 10^{-11})} + 1}$$

$$= \frac{0.03}{2033} = 1.48 \times 10^{-5} \ M \ CO_3^{-2}.$$

Using the activity product constant of 4.68×10^{-9} for calcite (5)

$$\frac{4.68 \times 10^{-9}}{1.48 \times 10^{-5}} = 3.16 \times 10^{-4} \ M \ Ca^{+2}.$$

At pH 12 and 25° C,

$$[CO_3^{-2}] = \frac{3 \times 10^{-2}}{\frac{(1 \times 10^{-24})}{(6.02 \times 10^{-11})(4.46 \times 10^{-7})} + \frac{(1 \times 10^{-12})}{(6.02 \times 19^{-11})} + 1}$$

$$= \frac{0.03}{1.0166} = 2.95 \times 10^{-2}\ M\ CO_3^{-2},$$

and

$$\frac{4.68 \times 10^{-9}}{2.95 \times 10^{-2}} = 1.59 \times 10^{-7}\ M\ Ca^{+2}.$$

Figure 10 is a graph of these data. The activity product constants used here (6) are evidently not applicable to these systems or the curve for fluorite would not overlap that of calcite. The general relationships are valid, however. Apatite could not be included because of uncertain composition (1, 6), and therefore solubility data. However, apatite does replace calcite in preference to fluorite at higher pH. We may conclude that the curve for apatite falls below that of fluorite above pH 7. The data of Table 1 support the premise that only one reaction product is stable in each system. In replacement reactions, a change in the resulting product is due to a change in the influent solution, all other things being equal. More than one final product indicates that the reaction has not been completed, or the original column contained an admixed solid that was less soluble in the system than

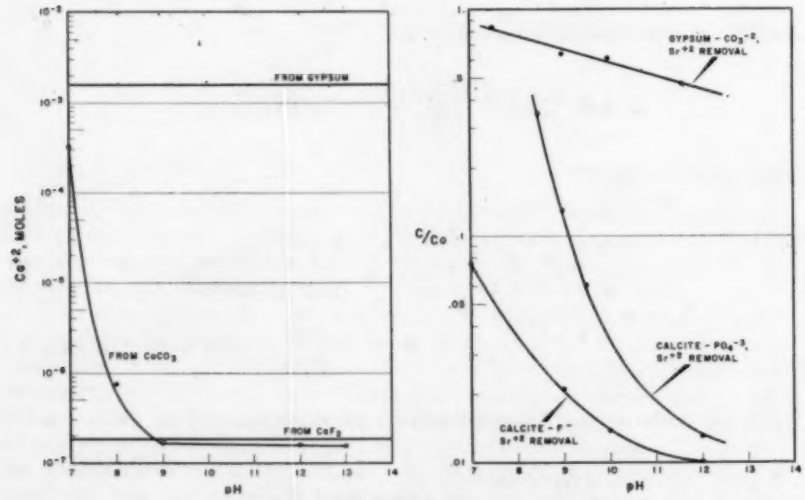

FIG. 10.

any possible reaction product if the influent solution has not changed in composition and pH. A column containing admixed calcite and quartz, for example, would yield apatite and quartz with an alkaline, PO_4^{-3}-bearing influent solution.

In the geological environment, there is the probability that a solution will pass through more than one type of wallrock, each of which may give rise to a different final product.

All attempts to approach the activity product ratio (4) experimentally from either direction have failed. Replacement reaction rates are exceedingly slow as the ratio is approached from either direction, and are zero at this ratio. While these activity ratios are of academic interest, they are nearly impossible to apply or check with any degree of accuracy. For example, a 1/1 mixed NaF-Na_3PO_4 alkaline influent reacts with calcite to form an apatite replacement product. As the PO_4^{-3} concentration or activity in the influent becomes progressively less, or approaches the ratio for apatite/fluorite, the calcite-apatite reaction becomes progressively slower until the calcite approaches stability in respect to the influent solution. The calcite-fluorite replacement reaction never takes place despite the F^- content of the solution. Thus, we have an essentially stable calcite-apatite-solution system from which one could hardly deduce the presence of a large fluoride concentration. In a single-cation system containing Ca^{+2} and several anions, only a single Ca^{+2}-anion compound is stable, and that compound is the one least soluble in the system. To have calcite, fluorite and anhydrite precipitating from the same solution, as Barton has suggested (5), is an impossible situation. One does not precipitate three compounds with a common cation (Ca^{+2}) at the same time, from the same solution. Even to have calcite and fluorite in equilibrium with the same solution implies that the activity ratios for F^- and CO_3^{-2} have changed; i.e., the solution was formerly either replacing fluorite with calcite or calcite with fluorite, and either the F^- or CO_3^{-2} activity decreased until the replacement reaction rate approached zero.

Activity ratios are a useful concept in the study of metasomatism, but do not constitute a substitute for detailed study of the reactions involved in metasomatism. Information about the composition of the solutions causing the metasomatism that can be deduced from a knowledge of activity product ratios is quite limited.

Figure 11 shows the removal of Sr^{+2} during the replacement reactions of Figure 9. Note that unlike the anion removals in Figure 9, Sr^{+2} removal always increases with increasing pH, probably due to a tendency toward decreasing hydroxide solubility. Ce^{+3} also exhibited increased removal with increased pH, suggesting that the phenomenon is common to all cations removed during anion replacement reactions.

The effect of the crystal structure of the final product on Sr^{+2} removal is even more evident. The crystal lattices of SrF_2 and CaF_2 are both of the fluorite type, while Ca-apatite and Sr-apatite are also of the same structure types. $SrCO_3$, on the other hand, is an aragonite type structure, while calcite is not. This structure difference inhibits the entrapment of Sr^{+2} in Ca^{+2} positions during the gypsum-calcite alteration by CO_3^{-2}-bearing solutions.

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CHEMICAL CHARACTERISTICS OF THE GROUND WATER RESOURCES OF JAMAICA, W.I.

1. A CONTRIBUTION TO THE INTERPRETATION OF HYDROCHEMICAL DATA[1]

V. G. HILL AND A. C. ELLINGTON

ABSTRACT

Frequency distribution methods are used to show that the ground water of the Clarendon Plains is not homogeneous and can be divided into three principal zones—the northern limestone belt, the central trough, and the coastal sand belt. The central trough type water can be further subdivided into sand and limestone types demonstrating the influence of lithology on the water. The specific conductivity, sodium and chloride values, are the parameters used to determine the frequency distributions.

INTRODUCTION

DURING the investigation of the hydrochemistry of the ground water of the Clarendon Plains there were indications that this water was not chemically homogeneous and so attempts were made to determine the pattern of the observed variations. The approach was to study statistically the distribution of some of the parameters measured, determine whether they were from a homogeneous population, and if heterogeneous indicate the groups present. This technique was later tested in other areas. The field work for this investigation was done during the period November 1959 and February 1960.

The Clarendon Plains is certainly the best developed and currently economically the most important aquifer in the island, so it was selected as the starting point in the present study of the ground water resources. It is hoped that the results obtained and the ideas developed can be applied to the lesser known aquifers thereby aiding in their successful exploitation. The programme of work includes a fairly detailed chemical analysis of the water samples for the determination of variations in composition. However, it is desirable to correlate these results with a parameter that can be easily measured. The specific conductivity of the water meets these requirements although its relationship to changes in chemical composition in a polycomponent system is complex.

The hydrology of the Clarendon plains has been studied in some detail by Stockley (5), Sweeting (6), Taylor (7), Versey (9), and Versey and Prescott (10). Versey (8) and Zans (11), have described the White Limestone, which is the principal aquifer in the region.

[1] Publication authorized by the Executive Secretary, Scientific Research Committee, Kingston, Jamaica W. I.

LOCATION AND GEOLOGY

The area comprises the coastal plain of the parish of Clarendon on the south coast and extends from the foot of the Mocho Mountains to the sea. Figure 1 shows the area and location of the wells. The relief is gentle rising gradually to about 300 feet in 8 miles. The Rio Minho and the Milk River

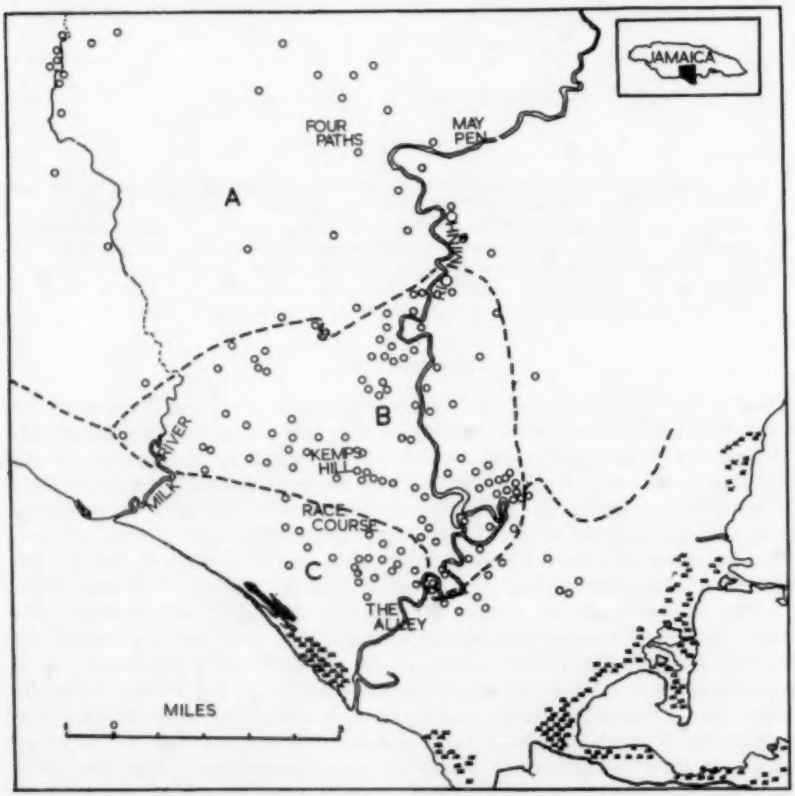

FIG. 1. Map of the Clarendon plains showing location of the wells, and the zones in the aquifer. A. Northern limestone belt; B. Central trough; C. Coastal sand belt.

are the two principal rivers draining the plains. Both are seasonal in flow, and are sometimes dry in places during the dry season. Numerous gullies that are usually dry serve to carry off a large part of the run-off water during the rainy season.

A blanket of alluvium that has a maximum thickness of over 750 feet overlies most of the white limestone formation in the region but low hills

of it project through this cover (9, 11). Of these Kemps Hill is the most prominent. According to Versey (9) the limestone appears to be compact but has a broken up rock head in the contact zone with the alluvium. He also suggests that this layer is the main aquifer. In this formation a fault occurs in an east-west direction along a line running from the vicinity of the Milk River Bath through to Kemps Hill.

The alluvium varies in composition from a transported bauxitic laterite in the Toll Gate area to a mixture of kaolinite and montmorillinite with sand and well rounded pebbles. The exchange ions on these clays are principally calcium and sodium. The sand and clay belt, and the underlying limestone are the two major aquifers recognized to date on the basis of lithology (5, 10, 11). Most of the wells in the limestone are artesian, and the piezometric surface varies from below five feet to over two hundred and fifty feet in the north western corner of the area. The depth of these wells varies from between eighty to six hundred feet. The ground water in the alluvium is partly under artesian conditions, and partly under water table conditions. These wells are between forty and four hundred feet deep. These depths are all measured from surface level (10).

SAMPLING AND CHEMICAL ANALYSIS

Samples from the wells are obtained from the discharge of the pumps at distances varying between six and twenty-five feet. In every case the sample was taken at the closest point possible, and the temperature and pH of the discharge were measured during sampling. The latter was obtained with B.D.H. narrow range indicator paper. If on arrival the pump was idle it was started and allowed to pump for from five to ten minutes before the sample was taken. Duplicate samples taken at the five and twenty minute interval after the pump was started gave results within the limits of the analytical error.

The conductivity of the sample was measured in the laboratory with a Mullard conductivity Bridge Type E7566 and the values corrected to 25° C. The sodium content was determined with an Eel flame photometer. The other constituents, except sulfate, were determined by the methods outlined in the "Approved Methods for the Physical and Chemical Examination of Water" (1). Calcium was determined by a versene titration at pH 12.5 using a sodium hydroxide buffer, while the combined calcium and magnesium were titrated at pH 9.5 using an ammonium chloride-ammonia buffer. The difference between these readings was taken as the magnesium equivalent (3). In some instances there was a rapid fading of the color of the Eriochrome Black T indicator at the end point of the combined calcium and magnesium titration. This was successfully overcome by adding a few crystals of hydroxylamine hydrochloride followed by potassium cyanide before titrating. The sulfate content was determined by weakly acidifying an aliquot of the sample with dilute hydrochloric acid, boiling for 10 minutes, adding 5.0 ml. of 0.020 N barium chloride solution, digesting for 5 to 20 minutes and allowing to cool. The whole was further cooled in a refrigerator before titrating

with versene. The ammonium chloride-ammonia buffer was used. A blank using an equal aliquot of distilled water was run at the same time. The sulfate equivalent for the standard versene solution used was calculated from the formula:

$$S = (B + T - R)\,F$$

where S = Sulfate content in parts per million, B = Burette reading for blank, T = Burette reading for the calcium plus magnesium titration, R = Burette reading for the sulfate titration, and F = Factor to convert values to parts per million sulfate. This factor takes into account the equivalence of barium and sulfate ions. An aliquot of 25.0 ml. sample was used in all the above titrations. A one-third excess of barium chloride is required, and the method fails if the sulfate content is too high. The values for the inorganic

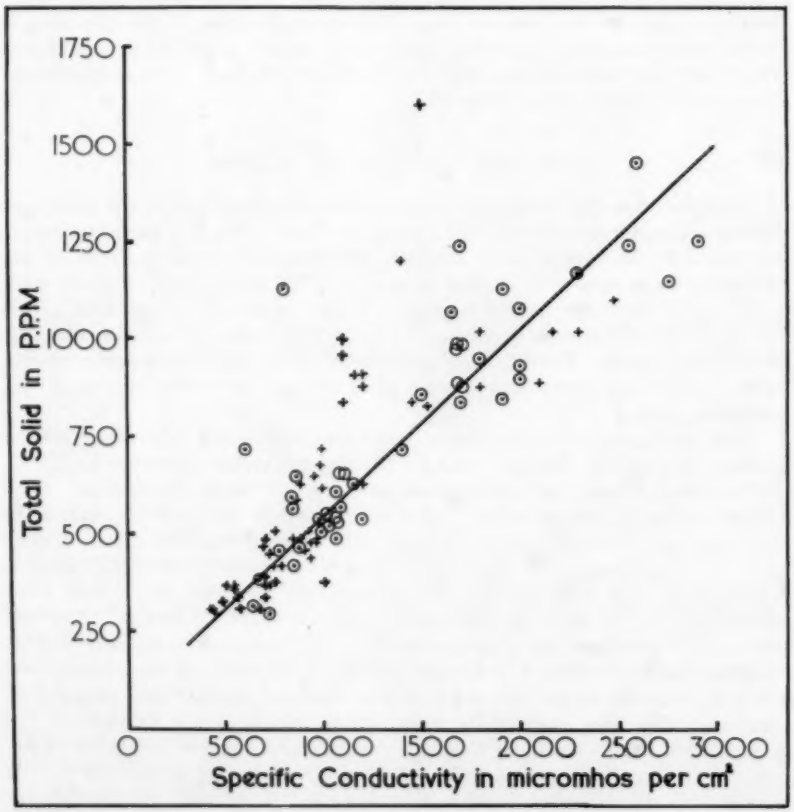

FIG. 2. Relationship between specific conductivity and total solids; + Samples from limestone wells; ○ Samples from sand wells.

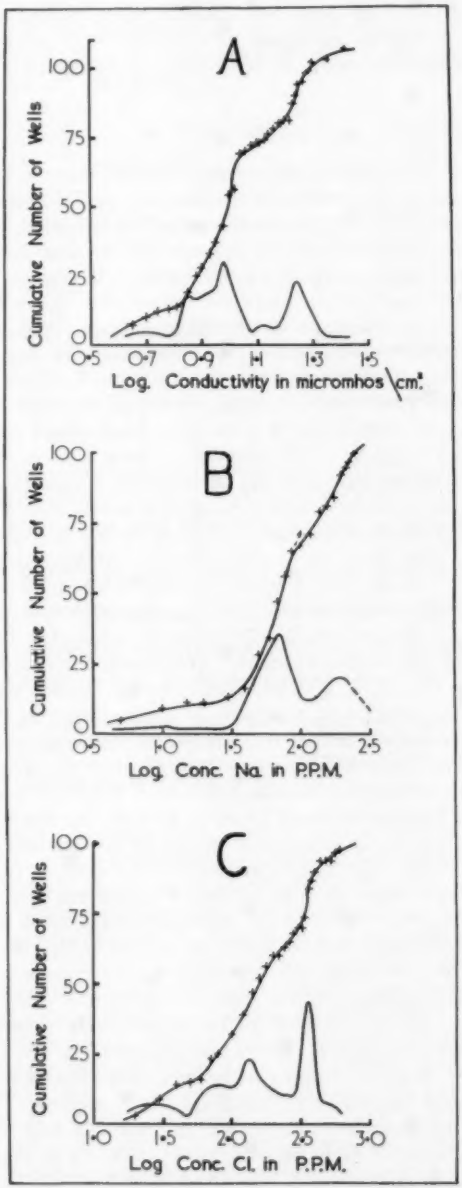

FIG. 3. Cumulative frequency distribution curves and derived unique frequency curves; A. specific conductivity; B. sodium; C. chloride.

constituents expressed in parts per million were added after converting the bicarbonate values to carbonate to give the total solids content. This avoids problems on the state of hydration of compounds formed during evaporation (4).

RESULTS

The results of the analysis are summarized in Figures 2 and 3. Figure 2 shows the relationship between the conductivity at 25° C and the computed total solids. The values for the sand wells are indicated by a cross (+) and those for the limestone wells by a circle (○). The distinction between sand and limestone wells is made on the basis of the horizon that produces the water. A statistical evaluation of the data shows that the regression line for the sand wells is not significantly different from those of the limestone wells. It is significant that the lines of regression for these two lithologic types of aquifers are similar in spite of the difference in ionic species present.

The above relationship between total solids and conductivity suggests that this value can be used as a single parameter to describe the sample. It is however important to know if the samples are from a homogeneous or mixed population and whether the observed distribution represents real or chance variations. It must be conceded that the sample points (wells) are not evenly spaced throughout the aquifer, but the fact that the wells were located by dousing suggests that they can be regarded as random samples from the region. Figure 3 shows the frequency distribution plot for conductivity, sodium and chloride concentrations. These parameters were selected because they do not have a marked affinity for the clays present. This effect would mask major trends in the distribution. Potassium, for example, tends to be strongly fixed by the clays, and so its distribution cannot be used for diagnostic purposes. It was found necessary to apply a log transformation to the concentration values of the parameters measured to normalize the distribution. A similar technique is standard practice in sedimentary petrography for grain size studies. The cumulative distribution curves obtained were differentiated by the method of Brotherhood and Griffiths (2) to give the unique frequency curve. They discussed the effects and errors in the smoothing of the cumulative curves, and the log transformation on the unique frequency curves. Close class intervals were used to minimise significant errors and spurious results. This technique removes the influence of the class interval used on the distribution obtained. All these curves are basically similar. In Figure 3, the A and C unique frequency curves show four modes while B shows only three distinct modes. When the cut off points 600, 810 and 1,330 micromhos per cm square are used for the conductivity values there is a distinct relationship in the distribution of the values (Fig. 1). The two outer peaks in Figure 3 A and C, appear to be related to the rock from which the water was produced, i.e. whether coastal sand or limestone. This association is supported by the boundary line shown in Figure 2, and suggests that the central peak in Figure 3 B is perhaps two peaks. The asymmetry of this peak supports this idea.

Table 1 shows the average composition of the different zones when there is

little or no pumping of the reservoir. Pumping, especially if excessive, produces changes in the composition of the reservoir, but the degree of that has not been ascertained with certainty. Work on this is still in progress and will be reported in the future.

DISCUSSION OF RESULTS

If the unique frequency curve is unimodal then it is reasonable to deduce that the samples are from a homogeneous population and are acted upon by the same set of conditions. However a bimodal distribution is indicative of two populations or possibly the effect of different factors resulting in what is essentially two populations. The presence of a polymodal distribution is

Table 1. Average composition of the water from the Clarendon Plains irrigation wells.

Zone	N[1]	°C	pH	Cond[2]	Ca[3]	Mg	Na	K	Cl	HCO_3	SO_4	SiO_2	PO_4	Total Solids
A	13	24.1	6.8	483	77.7	13.3	11.4	0.5	29.9	293.7	46.6	3.2	0.5	332.5
B	64	25.9	6.8	1000	105.8	23.9	93.3	2.6	175.5	317.9	38.4	5.9	1.1	608.3
B-l	36	24.4	6.6	1014	95.4	28.5	106.4	2.9	200.0	318.6	39.4	5.5	0.9	640.2
B-s	28	26.3	6.8	958	113.9	22.8	74.2	2.2	148.2	308.6	36.2	6.4	1.5	562.0
C	15	27.0	6.8	1750	185.9	37.2	129.5	2.7	337.6	353.3	50.5	3.3	1.4	927.5

A. Northern limestone belt. B. Central trough. B-l.Limestone zone of Central trough. B-s.Sand filled zone of central trough. C. Coastal "sand" belt. (1) Number of wells in group. (2) Values in micromhos per cm^2. (3) Values of ions in parts per million.

evidence that the samples are from mixed populations, or result from the effect of different and complex factors such as difference in the petrography of the reservoir acting on the sub-surface water. It is therefore necessary to compare all the factors and use different parameters to deduce information about the material being sampled. It is for this reason that three parameters, specific conductivity, sodium, and chloride ions were studied. The close similarity in the frequency distribution of the three parameters plotted suggests that there is a genetic reason for the determined distribution. The high and low modes are associated with the coastal "sand" zone and the "northern limestone" strip respectively (Fig. 1). The former is marked by a high conductivity, total solids, sodium and chloride values, while the latter zone has a low conductivity and total solids, with calcium and bicarbonate as the principal ions. The intermediate zone with a thick alluvium blanket is related to the two central peaks and is a distinct feature, probably due to the interaction of the aquifer and the contained water. The border between these

zones is not static, and shifts somewhat as the aquifer is pumped, but it appears that this shift is only within well-defined limits. The results suggest that some of the changes observed can be due to infiltration by sea water. When the wells are actively pumped equilibrium is not reached especially in the coastal "sand" belt and along the Kemps Hill fault. The Kemps Hill wells in the central trough zone are anomalous, and it is probable that along the fault there is movement of ground water admixed with sea water when the wells are actively pumped. This is probably the controlling factor. More work is being done on this phase of the problem as the interaction of sea water and other types of water with the aquifer is an important factor in ground water hydrology in Jamaica.

It is interesting that this attempt at a natural zoning of the aquifer should produce provinces that are genetically different. The presence of this distinct coastal "sand" type is important in that it delineates an area where the water is high in mineral content. In fact, the mean value is well above the recommended value for domestic use and it increases towards the sea. The salinity is particularly high and is well above the limit for agricultural use about one mile below the boundary. Water from this zone should be used with caution and only when it is not economic to get it from the central zone. In fact, it is not advisable that new wells be drilled very far below this line because of the possibility of overpumping the zone and so producing very saline water. A similar problem exists along the Kemps Hill fault.

The presence of a distinct "central" trough type water zone in the area is one of the novel results of this investigation. The influence of the clay minerals present on the composition of the water in the aquifer will be studied in detail, because the location and extent of the zone appears to be related to the buried Rio Minho valley. The proper solution of this problem is important because it also has direct bearing on the exploitation of other aquifers where a similar problem may exist.

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TEMPERATURES OF MINERALIZATION BY LIQUID INCLUSIONS, CAVE-IN-ROCK FLUORSPAR DISTRICT, ILLINOIS

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ABSTRACT

Temperatures of vapor bubble disappearance of liquid inclusions in fluorite, sphalerite, and quartz from the Deardorff and Victory Mines of the Cave-in-Rock fluorspar district were determined. Inclusions were studied in minerals of both principal stages of deposition indicated in the deposits. The data obtained indicate that ore deposition took place in the temperature range of 94° C to 142° C, assuming a pressure correction of +13° C, and that temperature oscillated during mineralization. Fluorite, the first mineral to be deposited in each stage, was deposited in the upper part of the temperature range, and sphalerite and quartz, which followed fluorite in each stage, were deposited at lower temperatures. Petroleum inclusions in fluorite show considerably lower temperatures of vapor bubble disappearance than aqueous inclusions, but petroleum inclusions cannot be used for temperature determinations, because pressure corrections cannot be estimated. The basic assumptions of the method are evaluated with respect to the deposits and it is concluded that no errors of first order magnitude are expected. The apparent variation in temperature during mineralization is tentatively interpreted as the result of two pulsations of hydrothermal solutions from a single source.

INTRODUCTION

THE liquid inclusion method of geologic thermometry has been used to determine temperatures of formation of many common ore minerals (12). Most studies have dealt with individual mineral specimens from various deposits; only a few systematic studies of fluid inclusions in representative suites of materials from mineral deposits have been made. In some cases, these studies have not given due consideration for the fundamental assumptions of the method. The present study deals with fluid inclusions in minerals from the Victory and Deardorff Mines, Cave-in-Rock district, Illinois. Its purpose is (1) to determine whether the fluid inclusion method is applicable to the minerals of the fluorspar deposits, (2) to determine variations in temperature during mineralization as indicated by the fluid inclusions, and (3) to evaluate the possible role of temperature variations in the genesis of the deposits.

The Cave-in-Rock fluorspar deposits are especially suitable for such an investigation for several reasons. First, the deposits belong to a group thought to have been formed at moderate temperatures and relatively shallow depths. In general, the assumptions of the inclusion method are less difficult to justify for deposits forming under these conditions. Second, the Cave-in-

Rock deposits have a relatively simple mineralogy, containing transparent or translucent minerals that generally have good crystal form. Third, the paragenetic sequence of the minerals is well established, and the stages of deposition are often readily apparent, even in hand specimens.

PREVIOUS WORK

The first descriptions of the geology of the Cave-in-Rock fluorspar deposits appeared in geologic reports on the nearby vein deposits of Rosiclare, Illinois. The descriptions were brief, for large-scale mining had not begun in the area. Detailed geologic descriptions of the bedded deposits appeared later in county reports of the Illinois Geologic Survey (2, 15). Papers dealing specifically with the origin of the deposits were published by Schwerin (14) and Currier (5). Results of investigations by the U. S. Geological Survey, U. S. Bureau of Mines, and the Illinois Geological Survey during World War II have been published in more recent years (6, 16). Some of these works give complete descriptions of the geology of the deposits and include discussions on the origin of the banding and comb structure. The various structures caused by replacement of limestone by fluorite have been described by Grogan (9). The reader should refer to the foregoing works for more comprehensive information than can be given here.

Fluid inclusions in fluorite from the Cave-in-Rock district were first studied by Grogan and Shrode (10). In 6 specimens of fluorite examined they found primary inclusions in zones parallel to cube faces and secondary inclusions along octahedral and other fracture planes. Which generation of fluorite they studied was not stated. The primary inclusions contained yellow petroleum, a vapor bubble, and some small globules of bitumen. Temperature values ranging from 83° C to 115° C were obtained. In three crystals, systematic variation in temperatures of vapor bubble disappearance from zone to zone was found. In each case, the highest temperature was given by an inclusion in the central zone (core) of the crystal. The secondary inclusions consisted of aqueous inclusions aligned along octahedral and other fracture planes. These gave temperatures of vapor disappearance ranging from 112° C to 172° C.

DESCRIPTIONS OF THE DEPOSITS

The Cave-in-Rock district in Hardin County, Illinois, is bounded on the south and east by the Ohio River and on the northwest by the Peters Creek fault zone. The district is underlain by Mississippian strata that dip to the northeast and are part of a large faulted structural dome. The deposits are typical bedding-replacement deposits that occur in the various favorable horizons of the Renault and Ste. Genevieve formations, as follows: (1) The Spar Mountain sandy bed (lower part of the Fredonia limestone member), (2) Various higher strata in the Fredonia limestone member, (3) The upper part of the Levias limestone and the basal part of the Renault limestone, (4) The uppermost part of the Renault limestone.

The fluorspar deposits are typically elongate and follow groups of joint-

like fractures or steeply dipping faults of small horizontal displacement (Fig. 1). They are cavity fillings and bedding-replacements of limestone. For the most part, the ore occurs in individual layers a few inches to several feet in thickness which are roughly parallel to the major bedding. There are commonly several mineralized beds separated by poorly mineralized or barren beds. The beds consist of bands of coarsely crystalline fluorite intercalated with bands of other minerals or unreplaced limestone. The fluorspar in the coarse layers consists of massive crusts of crystals, some of which face into open cavities and form a typical comb structure. Ore possessing the pronounced banding is locally termed "coontail ore." In the highly min-

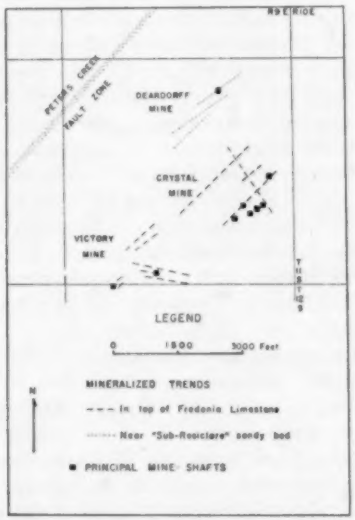

FIG. 1. Map of a part of the Cave-in-Rock district, Illinois (from Weller, et al.).

eralized rock, irregular cavities lined with crystals are common. Many of these cavities are elongate and aligned roughly parallel to the bedding. The aggregate thickness of the ore in the large deposits ranges from 4 to 15 feet. Toward the margins of the deposits, the number of mineralized beds decreases and the ore bodies pinch out.

Specimens for liquid inclusion study were collected from the Deardorff and Victory Mines and were taken at wide intervals along the strike of each of the ore bodies. In the Deardorff Mine, large amounts of limestone near the Spar Mountain sandy bed have been replaced by fine-grained sphalerite. Some of the coontail ore in this mine consists of alternating layers of coarse-grained fluorite and fine-grained sphalerite. The fluorite in the coontail ore is partly of the light colored early type and partly of the purple late type. Many open cavities in the sphalerite-fluorite coontail ore are lined with

well-formed crystals of sphalerite. Samples of both the massive coontail ore and the good crystals were taken, but temperature determinations were made only on the latter.

The sequence of mineral deposition in the Deardorff Mine is given in Figure 2, as determined by Weller, et al. (16). The sequence for the Victory Mine is approximately the same, though sulfides and quartz are uncommon. Two generations of fluorite have been recognized as well as two generations of sphalerite and quartz. Sphalerite and quartz deposition followed both periods of fluorite deposition. Calcite accompanied the later stage of sphalerite and quartz deposition. Very small amounts of chalcopyrite were deposited with fluorite. Pyrite, which is a very minor component, was deposited during and after each stage of fluorite deposition. Barite was apparently the last mineral to crystallize.

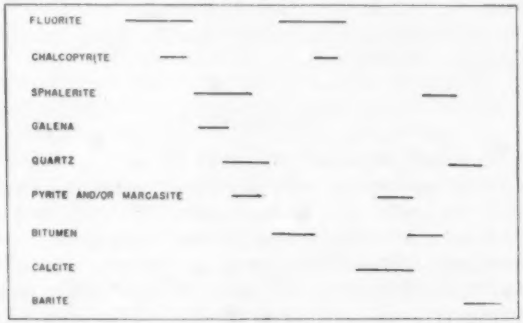

FIG. 2. Relative order of deposition in the Deardorff deposit (from Weller, et al.).

The early fluorite occurs mostly in the coarsely crystalline, light colored bands of the coontail ore, and occurs less commonly as crystals lining cavities. Much of the early fluorite is fractured, suggesting that the early-formed crystals were subjected to stresses, possibly caused by slumping. Single crystals large enough to section and not fractured are rare. Most of the fluorite of the Deardorff Mine apparently belongs to the early generation.

Fluorite of the late generation is typically purple or blue. It occurs both in the coontail ore and as clusters of crystals lining open cavities, but occurs most commonly as large crystals in cavities. The fluorite in the coontail ore is generally fractured and without good crystal faces, and it is not especially suitable for inclusion study. Some cavities in the early fluorite of the coontail ore are filled with purple fluorite. Late fluorite crystals also form growths of purple fluorite on colorless fluorite, which is presumably early.

Quartz and sphalerite vary greatly in abundance throughout the Deardorff Mine but occur in large amounts locally. The quartz occurs as tiny doubly terminated crystals encrusting other minerals in cavities. The late generation of quartz is subordinate in amount, and no specimens were obtained.

Both generations occur with sphalerite and the association suggests nearly simultaneous deposition of the two minerals. Globules of bitumen are common on the crystal surfaces and in some of the massive quartz layers. The sphalerite occurs as fine-grained layers in the coontail ore, or less commonly as small to medium-sized dark brown to black crystals in cavities. Most of the crystals are highly twinned. Calcite occurs as small to large milky or yellowish-brown crystals lining cavities. Many of these crystals are well-formed scalenohedrons.

Sphalerite and quartz are almost totally absent in the Victory Mine. Small disseminated crystals of quartz occur in some of the coontail ore, but no large concentrations have been found. The coontail ore of this mine is similar to that of the Deardorff Mine except that the ore contains very little sphalerite. In general, the occurrence and characteristics of fluorite are the same in the Victory Mine as in the Deardorff Mine. Some of the early fluorite, however, is yellowish rather than clear. Much of the calcite in the Victory Mine occurs as small milky crystals lining cavities or massive calcite in veins a few inches wide.

PROCEDURE

Samples from widely separated points in the mines were chosen for inclusion study. The specimens were sectioned with a diamond saw and mounted on Bakelite blocks with Duco cement. The mounted sections were ground to a thickness of about 1 mm on laps with various grades of carborundum abrasives. To prevent fracturing, sections of sphalerite were ground by hand to a thickness of 0.5 mm. A high polish was obtained on cloth laps impregnated with diamond powder. This step was omitted in preparing sections of fluorite and quartz. Instead, a thin coat of plastic was applied to each section to insure clear vision into the section. Following the grinding and polishing the sections were removed from the Bakelite blocks by soaking in acetone.

The heating apparatus is essentially the same one used by Bailey and Cameron (1). It consists of an aluminum ring mounted on an asbestos board ring with a sheet mica window covering the bottom. A circular copper plate with a small slit is placed over the aluminum ring, and a mica sheet and asbestos board flange are fitted over the entire cell. A Variac resistance coil is used to regulate the current to a nichrome heating coil surrounding the aluminum ring and insulated from it by a strip of sheet asbestos. The entire heating assembly is fixed in the jaws of a mechanical stage mounted on the stage of the microscope. The rate of heating that seems to give the most consistent temperature of disappearance of the vapor bubble of a single inclusion is 6° to 8° per minute. The temperature within the cell is recorded with a thermometer inserted through the side of the cell with the bulb directly under the crystal plate. The thermometer was calibrated with thermocouples inserted through the upper and lower mica windows.

Each section was studied carefully and those inclusions thought to be primary were selected for heating. The inclusions were observed during

heating with transmitted light and the temperature of disappearance of the vapor bubble of each was recorded. The temperature reading of each inclusion could be duplicated to within 1° C. In many cases, groups of primary inclusions could be observed and the temperature of vapor bubble disappearance of all the inclusions could be recorded.

RESULTS

Types of Inclusions.—Both primary and secondary inclusions were found in all of the minerals. The distinction between primary and secondary inclusions could be made quite easily in fluorite. Primary growth features are common in this mineral, and the distribution of inclusions with respect to these features is fairly obvious. Inclusions aligned parallel to color zones or in rows parallel to cube faces are considered primary. Irregular lines of inclusions that cross other lines of inclusions, or lines of inclusions parallel to the cleavage, are considered secondary. Primary inclusions in sphalerite and quartz are more difficult to recognize. Growth lines and color zonation are less common in these minerals, and even when these features are present, the distribution of inclusions with respect to them is not always apparent. Nearly all of the inclusions used in sphalerite occupy negative crystals, which are considered to be primary. Temperatures of vapor bubble disappearance for inclusions in the negative crystals are significantly different from those of obviously secondary inclusions.

Temperatures of vapor bubble disappearance of the secondary inclusions are consistently higher and, in some cases, 40° higher than those of primary inclusions. Secondary inclusions aligned along cleavage traces or fracture lines commonly give inconsistent temperatures. A large part of this variation is undoubtedly due to leakage during heating, a phenomenon commonly observed in secondary inclusions. Secondary inclusions, of course, cannot be used as indicators of the temperature of formation.

Two types of primary inclusions were recognized, aqueous and petroliferous, each having distinctly different temperatures of vapor bubble disappearance. The aqueous inclusions contain a clear liquid and vapor bubble. Petroleum inclusions, on the other hand, contain an amber-colored liquid, vapor bubble, solid bitumen, and a clear liquid phase, presumably water. Temperatures of the petroleum inclusions are consistently lower than those of aqueous inclusions in the same specimen.

In fluorite, the primary aqueous inclusions are elliptical or irregular in outline, or are negative crystals. Irregular inclusions predominate, but in some sections all the primary inclusions are bounded by cube faces. Many of the inclusions are aligned with respect to some growth feature of the crystal. Color zonation was found to be common in fluorite, especially in the late fluorite, and many of the inclusions are aligned in rows parallel to this feature. Many of the negative crystals have their longest dimensions oriented perpendicular to the growth lines with which they are associated.

Primary aqueous inclusions in sphalerite are rare and when present are difficult to observe due to the opacity of the mineral. Dark shadows along

the peripheries of the inclusions, caused by internal reflections, also made observations of the shrinking bubble difficult. Most of the inclusions are negative crystals. The aqueous inclusions of quartz are predominantly irregular in outline. Only those aligned along growth lines were used for temperature determinations.

Primary petroleum inclusions are very abundant in certain specimens of fluorite, but are rare in sphalerite and quartz. The inclusions are typically elliptical or spherical in outline, although some irregular inclusions occur. Most of the petroleum inclusions occur in rows parallel to growth lines of crystals. The inclusions contain a yellowish liquid vapor bubble or small globules of bitumen. Even when the petroleum inclusions are heated to a temperature 30° higher than the temperature of the disappearance of the vapor bubble, the bitumen does not appear to dissolve. An immiscible, clear liquid, thought to be water, is found in a few of the petroleum inclusions, but no other immiscible phases occur.

Temperatures of Vapor Bubble Disappearance

Deardorff Mine.—The majority of temperature data are for the Deardorff Mine, which was more abundantly sampled and has a greater variety of minerals than the Victory Mine. The ranges of temperature obtained are summarized in Table I.

Several sections of fluorite showed a systematic variation in the temperature of vapor bubble disappearance from the inner to the outer color zones. The maximum variation was about 15° C. In most of the crystal sections, the highest values were given by inclusions in the inner color zones. In one section, however, the highest temperatures were given by inclusions in the middle color zones. Along a single row of inclusions the temperatures of filling are usually very consistent.

Victory Mine.—Data were obtained only for fluorite in the Victory Mine and are summarized in Table II. Several sections of fluorite showed a systematic variation in temperatures of vapor bubble disappearance as in the

TABLE I

SUMMARY OF TEMPERATURES OF VAPOR BUBBLE DISAPPEARANCE IN PRIMARY INCLUSIONS—DEARDORFF MINE (UNCORRECTED FOR PRESSURE)

Mineral	Temperature range ° C	Number of specimens	Number of inclusions heated
Fluorite I			
Aqueous inclusions	113–129	3	25
Petroleum inclusions	88–111	2	8
Sphalerite I	84–103	2	13
Quartz I	81–105	3	21
Fluorite II			
Aqueous inclusions	110–129	8	91
Petroleum inclusions	—	—	—
Sphalerite II	81–92	1	9

TABLE II

SUMMARY OF TEMPERATURES OF VAPOR BUBBLE DISAPPEARANCE IN PRIMARY INCLUSIONS—VICTORY MINE (UNCORRECTED FOR PRESSURE)

Mineral	Temperature range ° C	Number of specimens	Number of inclusions heated
Fluorite I			
Aqueous inclusions	114–121	1	6
Petroleum inclusions	—	—	—
Fluorite II			
Aqueous inclusions	114–128	3	34
Petroleum inclusions	70–83	2	36

Deardorff Mine. The temperature range for both generations of fluorite is essentially the same as that from the Deardorff Mine.

Data from both mines are combined in Table III. Note that temperatures of vapor bubble disappearance of petroleum inclusions are much lower than those of aqueous inclusions. The temperature range indicated by the petroleum inclusions, however, is within the range obtained by Grogan and Shrode (10), who also measured temperatures of vapor bubble disappearance of petroleum inclusions.

SOURCES OF ERROR

The use of liquid inclusions in geologic thermometry has received considerable attention; many geologists have questioned the validity of application of the method. Some have pointed out the limitations and sources of error in determining temperatures of formation (1, 12, 13). Each new investigation must include an appraisal of the validity of the basic assumptions of the inclusion method with respect to the geology of the deposits. The basic assumptions have been summarized by Bailey and Cameron (1, p. 642) and are listed as follows:

1. The liquid completely filled its cavity at the time of crystallization.
2. Primary liquid inclusions can be distinguished from secondary inclusions.

TABLE III

TEMPERATURES OF VAPOR BUBBLE DISAPPEARANCE IN PRIMARY INCLUSIONS—DEARDORFF AND VICTORY MINES COMBINED DATA (UNCORRECTED FOR PRESSURE)

Mineral	Inclusions type	Temperature range ° C
Fluorite I	Aqueous	113–129
	Petroleum	88–111
Sphalerite I	Aqueous	84– 92
Quartz I	Aqueous	81–105
Fluorite II	Aqueous	110–129
	Petroleum	70– 83
Sphalerite II	Aqueous	81– 92

3. There has been no significant change in the volume of the cavity itself due to pressure, solution or precipitation.
4. Representative samples are used for temperature determinations.
5. The pressure on the liquid at the time of inclusion was small, or its magnitude can be estimated.
6. The liquid is an aqueous solution containing no carbon dioxide or other gas in large concentration.
7. There has been no addition of liquid to the cavity or loss of liquid from it.

As pointed out by Bailey and Cameron, each of the above assumptions is a possible source of error, completely apart from the experimental errors. The temperature of disappearance of the vapor bubble may fail to approximate the temperature of formation, if any one of the above assumptions is not satisfied.

Some of the basic assumptions can be made for the Cave-in-Rock fluorspar deposits with considerable confidence. The probability that the cavities were filled with liquid at the time of inclusion is strong. The pressure at the depth at which the deposits probably formed (160 to 235 atmospheres) would be sufficient to keep water in the liquid state. The boiling point of pure water at this pressure is greater than 350° (11, p. 1927); therefore, it is likely that no water vapor was included in the cavities. Furthermore, available data on the composition of liquid inclusions in fluorite of this district indicate that the liquid is a saline solution. The boiling point of this solution would be higher than 350° C. The presence of volatile solutes would lower the boiling point, but very high concentrations would be required to lower the boiling point to the range of temperature within which the minerals are thought to crystallize.

The criteria used for distinguishing primary inclusions from secondary inclusions have been discussed in the previous section. It is thought that the "apparently primary" inclusions used in temperature determinations would be accepted as primary without question by at least the majority of students of liquid inclusions.

The assumption that there has been no significant change in volume of the cavity can be made with some assurance. Essentially the same arguments made by Bailey and Cameron (1) in this connection can be used here. The compressibilities and solubilities of the minerals studied are so low as to make very unlikely changes in volume of the cavity during heating. Precipitation of solute is also unlikely because the liquid is apparently undersaturated with respect to the principal solutes as indicated by the absence of visible solid phases in the liquid.

For temperature study, the necessity for using more than a single sample of each mineral from a deposit is readily apparent. In general, temperature data are most indicative of the temperature range of deposition when they are obtained from specimens taken from many parts of the deposits. In the present investigation, specimens of fluorite were collected along the strike of the main ore body in each mine. Sphalerite, quartz, and calcite were

collected only in areas of greatest abundance, these being in local areas of the Deardorff Mine. The collection made by the writer includes specimens that represent the typical occurrences of each mineral in each mine, and it is felt that the collection is a satisfactory sampling of the two ore bodies for the present purpose. Because it was not possible to obtain temperature data for all samples collected, including certain typical ones, the temperature data may not represent the full range of deposition. For example, the largest volume of early fluorite was deposited as coarsely crystalline coontail ore. No data for primary inclusions in this material were obtained because of difficulties in preparation and examination of sections. Only the well-formed crystals of early fluorite from cavities in the coontail ore could be used for temperature study, for these contained abundant large inclusions that could be easily recognized as primary. These crystals, though of the first generation, do not represent the earliest formed fluorite, and it is therefore possible that the determined temperature range does not cover the full range of temperature during deposition. On the other hand, specimens of late fluorite were taken from cavities in the highly mineralized rock, and because this is their typical occurrence, the temperature range is probably the full range.

Still another limitation on the reliability of the data is the number of specimens and inclusions in each specimen studied. Inclusions are far more abundant in the late fluorite than in any of the other minerals, and in sphalerite are exceedingly uncommon. The data, then, are far more reliable for the late fluorite than the other minerals if numbers of observations are a limiting factor. But it is felt that they are sufficiently reliable for all the minerals at least to indicate the general pattern of temperature variation during mineralization.

There is next the problem of estimating the original pressure. To do this, one must rely on estimations as to the depth of burial at the time of mineralization, which are admittedly not at all certain. Assuming that the depth at which mineralization took place was about 3,000 feet, as suggested by Weller, et al. (16), the original pressure must have been between 235 and 160 atmospheres, the calculated lithostatic and hydrostatic pressures respectively. A temperature correction of 10° to 15° C for these pressures should be added to the temperature readings (13).

The correction that must be made for each temperature reading due to the original pressure is calculated from the P-V-T relations of water. Nonvolatile solutes, such as NaCl, tend to reduce the compressibility of water and will affect the temperature of vapor bubble disappearance. A vacuole containing a salt solution, therefore, will have a higher temperature of vapor bubble disappearance than one containing pure water, and its pressure correction will be correspondingly lower. It is fairly certain that the liquid of the inclusions is not saturated with respect to salts. In the absence of quantitative data on the salt concentration of the inclusions, all that can be said is that the effect of solutes on the temperature readings will be small.

Data for petroliferous inclusions are unreliable, because the presence of hydrocarbons means that the pressure corrections based on P-V-T curves for water cannot be applied. Addition of hydrocarbons to water greatly in-

creases its compressibility and the temperatures of vapor bubble disappearance values in such inclusions will be much lower than those obtained from inclusions composed of pure water formed at the same pressure. This was observed in fluorite specimens that contain both aqueous and petroleum inclusions (Table III). The appropriate correction could be applied to data for petroleum inclusions if the composition of the petroleum solution were known and the P-V-T relations of that system were available. Without such data, however, petroleum-containing inclusions are unreliable indicators of temperature of mineral formation.

The assumption that the liquid contains no large concentration of gas will be discussed with reference to CO_2, the gas most likely to be present in the mineralizing fluids. The presence of CO_2 in the inclusions will not seriously affect the temperature of vapor bubble disappearance if the liquid is undersaturated with respect to CO_2. The actual amount of CO_2 that could dissolve in H_2O in the range of temperature and pressure of mineralization is low. Garrels and Richter (8) give the solubility of CO_2 in H_2O at 115° C and 200 atmospheres as about 2.2 mole percent, and the compressibility of water is not greatly changed by the presence of this amount of CO_2 (8). The temperature of vapor bubble disappearance of an inclusion containing small amounts of CO_2 will not be significantly different from an inclusion containing pure H_2O. Furthermore, the error caused by the presence of small amounts of CO_2 in solution would be well within the range of experimental error and the range of pressure correction that is applied to the temperature readings.

A serious error in the data may result if the liquid entrapped is supersaturated with respect to CO_2 and small amounts of gaseous CO_2 are entrapped with the liquid. The degree of filling will then be less than that of inclusions in which only liquid was entrapped. When these inclusions are reheated the temperature of disappearance of the vapor bubbles will be higher than the temperature of disappearance of the vapor bubble of inclusions which had no gas bubbles entrapped with liquid.

There is some indication that the liquid is not supersaturated with respect to CO_2. Inclusions along a single growth line in fluorite have very consistent degrees of filling. If gas bubbles were entrapped with liquid in the inclusions, the proportion of gas to liquid at the time of entrapment must have been the same to give a consistent degree of filling. Considering the probable erratic distribution of gas bubbles on the surface of the growing crystals, it is very unlikely that the same proportion of gas to liquid would be entrapped in each inclusion. It seems likely, therefore, that the liquid was undersaturated with respect to CO_2 and that no serious error from this source is to be expected.

The final assumption to be considered is that no liquid has been added to or subtracted from the cavity since its formation. Some of the inclusions, especially secondary inclusions with exceedingly large vapor bubbles, have obviously leaked. Leakage may have occurred along fractures, lineage boundaries, and cleavage surfaces where these features intersect lines of primary inclusions. Some inclusions leaked upon heating as indicated by a sudden

expansion of the vapor bubble. This type of leakage is easily recognized by careful observation, but other types of leakage, which may affect temperature readings systematically, are much more difficult to appraise.

Kennedy (13) has suggested that a pressure differential between the liquid in the cavities and the liquid surrounding the crystal may have occurred as the temperature declined after mineralization. Some of the liquid surrounding the crystal would migrate into the cavities in response to the pressure gradient. This would systematically lower the temperature of vapor bubble disappearance when the inclusion is reheated. He cites the results of a few experiments showing that a decrease in vacuole size is produced by application of water vapor at high pressure to samples of fluorite. The high pressure differential which was attained in these experiments is probably much greater than ever attained in the Cave-in-Rock deposits; hence the results of Kennedy's experiments probably have little bearing on the results of the present investigation.

In conclusion, evaluation of the basic assumptions indicates that temperatures of vapor bubble disappearance of inclusions in the minerals studied should give at least a first approximation of the temperature of mineral formation. Certain small systematic errors may be present, such as that due to the uncertainty of pressure corrections, but no errors of first-order magnitude are expected. Data are more reliable for the minerals containing an abundance of inclusions than others, but at least the general pattern of temperature variation during mineralization is indicated.

DISCUSSION

Several interpretations of the apparent oscillation in temperature during mineralization need to be considered. First, the temperatures of mineral formation as determined by liquid inclusions can be accepted as indicating the actual variation in temperature of the ore solutions as they arrived at the site of deposition. If this was the case, then the deposits were formed by two pulses of mineralizing fluids that entered the ore zone at about the same temperature and deposited first fluorite, then sphalerite and quartz with falling temperatures. The two emanations of mineralizing fluids must have been of similar composition, as well as at the same temperature, in order to deposit the same minerals in each stage. The source of the fluids must not have cooled appreciably between emanations since both stages of deposition occurred in the same temperature range. Possibly the mineralizing fluids left the source at closely spaced intervals so that sufficient time had not elapsed for loss of heat from the source.

If deposition in each stage of mineralization occurred with falling temperatures as is indicated, the drop in temperature may have been an important factor affecting deposition. No attempt is made here to discuss possible mechanisms of transportation and causes for deposition of the ore minerals in these deposits. However, it appears that from the evidence at hand, changes in temperature should be carefully considered in a discussion of the genesis of these deposits. Possibly the fluorite was formed by reaction

of HF with limestone as Currier (5) suggested. If so, then the metal ions of Zn, Cu, Fe may have been carried in the same solutions as fluoride complexes. These may have been broken down, freeing the metal ions to be deposited as sulfides, as the F-ion activity was decreased by reaction with limestone to form fluorite. Or the metal ions may have been carried as complexes (fluoride, chloride, or sulfide) and a drop in temperature may simply have caused the breakdown of these with consequent deposition of sulfides. No doubt, there were many reactions involved, and certainly other variables besides temperature were important.

A second possible interpretation of the apparent temperature variation involves the effect of exothermic reactions on the temperature of mineralization. The temperature of vapor bubble disappearance is considered to approximate the temperature (with pressure correction) of the inclusion at the time of its entrapment. This temperature may or may not be the same as the temperature of the incoming ore solutions if heat is exchanged during crystallization. For example, heat may be evolved during crystallization that could raise the temperature of the ore solutions immediately surrounding the crystal. Temperature values obtained from the liquid inclusions would therefore be higher than the temperatures of the incoming ore solutions. An oscillation of temperature during mineralization then could be explained if crystallization of certain minerals in the paragenetic sequence involved exothermic reactions, which raised the temperatures of deposition of these minerals above the temperatures of the ingressing ore-bearing solutions.

Crystallization of fluorite, if it occurred according to the reactions proposed by Currier (5), would evolve a certain amount of heat. On the other hand, crystallization of quartz and sphalerite, assuming that they are carried in ionic solution, would evolve little or no heat. Therefore, it is possible that the temperature of the solutions would be raised by the deposition of fluorite, so that fluorite would be deposited at higher temperatures than quartz or sphalerite. If this were the case, then the heat evolved must not have been transferred away from the site of deposition before entrapment of the liquid in the inclusions. Obviously heat losses from the site of deposition would depend on the rate of evolution of the heat and the rate of heat transfer from the ore zone. Both of these variables are unknown and indeterminate quantities.

Possible increase in temperature of mineralization by exothermic reactions has been reported in halite crystals by Dreyer, Garrels, and Howland (7) and discussed with reference to other minerals by Bailey and Cameron (1) and Cameron, Rowe, and Weis (3, 4). It has been pointed out that not enough is known about the reactions causing precipitation to evaluate the effects of exothermic reactions. Certainly this failing applies to the present investigation, for at the present time the nature of transportation and deposition of the ore constituents in the Cave-in-Rock deposits is very uncertain. It is very difficult to argue, however, on the basis of hypothetical reactions that exothermic reactions caused the apparent higher temperatures of fluorite deposition. In view of this and other uncertainties it would seem more reasonable and certainly simpler to accept for the time being the first ex-

planation: that the observed temperature variation during mineralization resulted from two pulses of mineralizing fluids emanating from a single heated source at depth.

SUMMARY

Two generations of minerals are recognized in the fluorspar deposits of the Cave-in-Rock district. Although there are some variations in abundance of minerals and in paragenetic sequence from mine to mine, the general order of deposition of the minerals studied is as follows: fluorite, sphalerite, quartz, fluorite, sphalerite, and quartz. Liquid inclusions in these minerals have been heated and the temperatures of vapor bubble disappearance observed visually. Primary and secondary inclusions have been studied and in all cases secondary inclusions give higher temperature values. Both aqueous and petroliferous primary inclusions have been observed. Temperatures of vapor bubble disappearance of petroleum inclusions are consistently lower than those of aqueous inclusions. The former cannot be used as indicators of the temperature of formation until a method of determining a pressure correction for such inclusions can be developed. Temperatures of vapor bubble disappearance of primary aqueous inclusions indicate that mineralization took place in the range of 94° C to 142° C. Fluorite was deposited in the range of 110° C to 129° C, whereas sphalerite and quartz were deposited in the range of 94° C to 118° C (assuming a pressure correction of + 13° C). Thus the temperature appears to have oscillated during mineralization. The apparent temperature variation is tentatively interpreted as the result of two pulsations of hydrothermal solutions from a single source. Effects of exothermic reactions on temperatures of mineral formation cannot be evaluated at the present time, for the nature of ore transportation and deposition in these deposits is too imperfectly known. The basic assumptions of the inclusion method are evaluated with respect to the deposits and it is concluded that the temperatures of vapor bubble disappearance give close approximations of the temperatures of mineral formation.

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RADIOACTIVITY AND TERTIARY VOLCANIC ACTIVITY IN EGYPT

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ABSTRACT

Tertiary tectonic disturbances in Egypt produced widespread basaltic fissure eruptions and diverse hydrothermal activities that include copper, iron, iron-manganese and lead-zinc mineralizations. Studies on the trace radioactivity of some of these deposits suggest the presence of a weak uraniferous phase in some or in certain stages of these hydrothermal activities, and can explain the known occurrences of economic uranium deposits in Egypt.

These volcanic hydrothermal solutions were not originally uraniferous but have become so during their ascent by leaching and mobilizing pre-existing labile uranium and other ions in the traversed rocks.

INTRODUCTION

THE late Tertiary epeirogenic movements in Egypt were accompanied by intensification of structural deformations that had already been recurrently initiated in the late Cretaceous. These deformations were mostly of tensional nature such as normal faults best exemplified by those responsible for the Red Sea graben and those near it. Simatic molten material had climbed up along abyssal fissures controlled by the more important structural lines of weakness and are represented now in Egypt by generally late Oligocene lava flows, sills, sheets, plugs, necks and dike-feeders, chiefly of olivine-basalt and olivine-diabase. Although the majority of these individual igneous bodies does not reach large dimensions (as compared for example with those of the neighboring Tertiary Syrian volcanic fields), they are rather well dispersed all over the country, somewhat increasing to the northeast. Andrew (2) had compiled the occurrences and field relations of the majority of these igneous bodies.

More important still for our purpose are the hydrothermal volcanic activities that had accompanied such magmatic eruptions and which had long survived them in Egypt. These may be even represented now by certain thermal springs of the country. Relics of fumaroles, gas maars (17), mud and sand volcanoes mostly alined along fissures and faults are known to occur. Rittmann (13) had theorized the different mechanisms by which the volcanic gaseous constituents could have separated from their basic magmas in order to explain the diversified Tertiary volcanic phenomena displayed in Egypt.

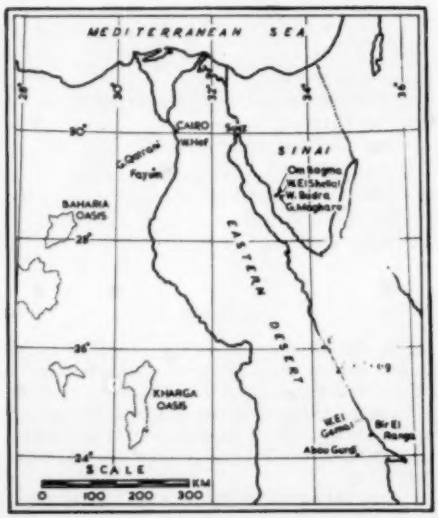

FIG. 1. Sketch-map of Egypt indicating positions of localities mentioned in the text.

The mineralizing effects of these hydrothermal activities include widespread ferruginous and manganiferous coloration of the arenites, and dolomitization, silicification or coarse recrystallization of limestones and chalks along fault and joint planes. Of economic importance are the metasomatic replacement of topmost Cretaceous and Eocene limestones and chalks of Baharia Oasis by goethite and hematite (9), of Carboniferous dolomites and dolomitic limestones of west central Sinai by Miocene manganese-iron oxides (3 and 10), of some Miocene sediments by lead and zinc, and of some copper ores filling fissures dispersed in Sinai and the Eastern Desert. There is now a justified, though mostly tacit, belief among geologists in Egypt that the bulk of these mineralizing ions was not originally indigenous to the Tertiary basic magmas or their vapors but represented ions leached and mobilized by the uprising fluids from the deeper Precambrian rocks and ore deposits and/or from their overlying sediments, which were permeated and

channeled by the volcanic hydrothermal fluids. Evidence of protracted or renewed hydrothermal activities and mineralizations along weak structural planes or porous rocks is also available.

In the course of recent studies carried out by the present writer on the radioactivity of several rocks and mineral-ores from different parts of Egypt, there emerged evidence suggesting a hitherto unsuspected relation between the traces of radioactivity of some of these Tertiary mineral deposits and the hydrothermal Tertiary volcanic activity. Uranium, like many of the Tertiary mineralizing ions was also leached and mobilized by the volcanic fluids. This evidence is given below and it is also shown that all economic uranium concentrations at present known in Egypt can also be related to the Tertiary volcanic hydrothermal activities. Laboratory studies of the radioactivity were carried largely by means of nuclear emulsion techniques on adequately stored polished samples and powdered spot or composite, rock or mineral, samples carefully split down to suitable sizes. All unspecified values of radioactivity mentioned below refer to the number of alpha particles, multiplied by 10^{-6}, emitted from a surface of one centimeter square of the sample in one second (alphas $\times 10^{-6}/cm^2/sec$). Localities mentioned in the text are indicated in Figure 1.

RADIOACTIVITY

Minor Tertiary Dolomite, Ferruginous and Siliceous Deposits.—Joint surfaces of the Turonian limestones of Abou Roash near Cairo are dolomitized to depths of six inches or more. Any helictites previously formed on the joint surfaces are likewise dolomitized. The dolomitized parts have a radioactivity of about 113 (eU_3O_8 1.3 ppm) but the limestones themselves are almost non-radioactive. Some dolomitized fault surfaces in limestones of the same area contain an eU_3O_8 of 2.5 ppm. The dolomitization has been considered as due to hydrothermal activities connected with the volcanic Tertiary igneous rocks of that area (8).

Impressive ferruginous tube and pipe-like structures in the Oligocene sandstone of Gebel Ahmar, east of Cairo, are considered to be due to rising fluids carrying iron, manganese and sulfur oxides, in the form of fumaroles, and are related to the dwindling stages of Tertiary volcanicity in the same area (16). Many of these ferruginous tubes are at least twice or thrice more radioactive than the normal sandstone. The same applies to ochreous colorations ascribed to a still later phase after the formation of the tubes, but the tenor of color is not proportional to that of the radioactivity.

These examples are probably the least satisfactory to demonstrate a relation between the radioactivity and the hydrothermal volcanic activities in rocks of both areas. This is because of the relatively low traces of radioactivity involved and because of interfering complicating factors including the effects of any circulating later meteoric or intrastratal solutions. Several of the silicified rocks, coarsely recrystallized calcite and iron-stained sandstones commonly attributed to Tertiary hydrothermal activities from both areas and other similar areas, e.g., the Wadi Hof area south of Cairo, and the Oligocene-Miocene belt between Cairo and Suez, do not show any ap-

preciable difference in their radioactivities from those of their original rocks. Some affected rocks may be even less radioactive than their parent rocks indicating leaching of part of their original radioelements by the volcanic thermal waters. Some of the silicified tree trunks in the Oligocene Sandstone between Cairo and Suez are appreciably radioactive. The radioactivity is concentrated in small reddish ferruginous clayey veinlets and in holes and microfractures traversing the silicified wood cells. The origin of this radioactivity is, however, debatable. The total radioactivity of the silicified trunks varies to an observed maximum of 10065 (eU_3O_8: 144 ppm).

Pyrite Replacements in Kharga Oasis.—In Kharga Oasis, casual large cubes of pyrite (sides of cubes from 1 to 2 cm) occur as metasomatic replacements in the early Tertiary limestones and chalks, particularly increasing in the vicinity of some joint and fault planes. The cubes are now completely pseudomorphed by goethite. They contain an average of about 103 ppm of U_3O_8 whereas their host rocks are almost non-radioactive. Goethite pseudomorphs after pre-existing aggregates of coarse rhombic carbonate (siderite?) similarly occur in the limestones and chalks but are less radioactive with U_3O_8 ranging from 11 to 25 ppm. The only originally somewhat radioactive beds in the stratigraphic succession there are in the underlying Maestrichtian "Lower Esna Shales," phosphate beds and in some lignitic shale bands in the basal Nubian Sandstone. The shales, especially the "Lower Esna Shales" are considered to have been deposited originally in lagoons of restricted aeration (black shale or euxenic facies of sedimentation) and they have a rich content of fossil flora (fruits, branches, etc.) and dwarfed fauna (11). There is good evidence of the pre-existence of much disseminated iron sulfides as well as iron-sulfide rosettes, concretions and pseudomorphs of fossils. In exposed outcrops all the iron sulfides are replaced by goethite though some are still preserved fresh in the cores of certain large fossils. The average radioactivity of the "Lower Esna Shales" there corresponds to an average content of about 25 ppm of U_3O_8 (determinations on exposed brownish surface samples but may be much higher in fresher core samples unavailable to the writer). It seems appropriate therefore to suggest that the ascending Tertiary thermal solutions had leached part of the iron sulfides and uranium from the Maestrichtian black shales and had deposited them in the overlying Tertiary limestones. Tertiary faulting is common in the area and to it the Oasis owes its origin.

Copper Deposits of West Central Sinai.—In several areas of the Sinai peninsula, joint, fault, and minor fissure surfaces of the Precambrian crystalline basement as well as the overlying Carboniferous to Upper Cretaceous sediments are encrusted by copper minerals, chiefly the carbonates, less commonly chrysocolla and turquois, and rare cuprite, chalcocite, and covellite. Fissure and fault fillings are also common but so far no important economic copper deposits have been discovered (6). On the basis of minor trace element analysis of the rare chalcocite (containing traces of Sn and Bi), the Sinai copper ores are considered to have been of an original mesothermal hypogene origin whereas the other copper minerals are later oxidation products from the chalcocite by meteoric waters (4, 6). Malachitic stains also

color some of the Carboniferous siltstones and shales and seeps for short distances along their bedding planes.

Malachite scraped from small fissure-fillings or from concretionary colloform growths lining open planes in the crystalline basement rocks of west central Sinai always shows a persistent average U_3O_8 content ranging from 40 to 60 ppm and thereby differs from malachite in the oxidation zones of copper ores of the classic sense in the Precambrian ores of the Eastern Desert. The latter is almost non-radioactive, e.g. malachite of the oxidation zone at Abou Gurdi, near latitude 24° contains U_3O_8 in the range of 1 to 0.23 ppm. One chrysocolla dike-like deposit in west central Sinai has a total radioactivity equivalent to a content of about 27 ppm. U_3O_8, and turquois encrustations on the joint surfaces of the Carboniferous Lower Sandstone in one of the historical mines of the area (Maghara) have an average U_3O_8 content of 212 ppm. In all these cases, host rocks of west central Sinai have their usual low radioactivity. Autoradiographs of polished specimens of colloform concretionary malachite from Sinai indicate that uranium is fairly evenly distributed within the bulk of the mineral and suggest coprecipitation of uranium and copper carbonate from solution.

From radioactivity and field evidence, it is reasonable to suggest that the malachite, turquois, and chrysocolla of west central Sinai were not entirely products of a straightforward meteoric oxidation of primary sulfides but must have partly at least formed through the participation of oxidizing hydrothermal uprising solutions that were carrying some uranium.

Manganese-Iron Deposits of West Central Sinai.—The productive and important manganese-iron deposits of west central Sinai occur in isolated lenses and lenticular or tabular bands selectively replacing the dolomite and dolomitic limestone intercalations of the Carboniferous formation through feeders along fault planes of Miocene age (3, 10). In certain localities, this dolomite reveals in thin section ghosts of a replaced pre-existing oolitic texture within the dolomite rhombs; suggesting that the rock is actually a dolomitized calc-arenite (pers. observation). This dolomitization is considered, however, to have been contemporaneous with deposition on the Carboniferous sea-shelf and at least is not related to the hydrothermal Tertiary activities. Tertiary diabasic sills, sheets, and dike feeders in fault and joint planes are common in west central Sinai.

El Shazly and Saleeb (7) have recently described properties of the different minerals in the Sinai manganese-iron deposits and elsewhere. In Sinai, the chief and abundant ore mineral is ferruginous pyrolusite in diverse habits: massive, concretionary, finely or coarsely crystalline in long (up to 10 cm) delicate subparallel, radiating or plumose and leaf-shaped aggregates, in stout acicular metallic-looking prisms, or in tiny interlocking decussate fibers.

The earliest stages of manganese-iron replacement of the Sinai dolomite appear in extremely narrow meandering or straight (along incipient joints) black streaks and spots of pyrolusite in the dolomite, (Fig. 2). These gradually widen and increase in dimension. The radioactivity of the replaced dolomite always progressively rises with the increase of this pyrolusitization.

Large ore samples from the different mines of west central Sinai display variable radioactivities equivalent to U_3O_8 contents from 1 to 140 ppm. While the factors controlling this variation in radioactivity of the ores have not been yet specifically studied, preliminary indications from several samples from mines of the different individual deposits suggest that each deposit tends to have a specific range of radioactivity. Manganese-iron ore samples of equal grades from different deposits may have quite different radioactivities. The most common range of radioactivity corresponds to U_3O_8 contents in

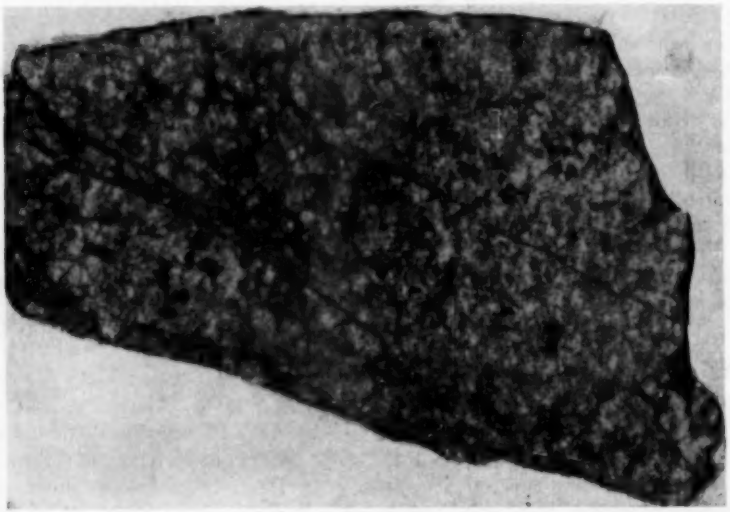

FIG. 2. A polished slab of dolomite showing incipient stages of replacement by black streaks and clots of microcrystalline fibrous pyrolusite. Average radioactivity of a powdered sample from this mineralized dolomite is 397×10^{-8} alphas/cm^2/sec (eU_3O_8 content of about 4.5 ppm) three to five times greater than that of the same non-mineralized dolomite. Om Bogma mines, west central Sinai, about twice natural size.

the range of 20 to 40 ppm. The least radioactive manganese ore samples hitherto observed all come from the lenses in the dolomite of Wadi Budra-Wadi el Shellal areas. Twelve composite samples of ores from these two areas have U_3O_8 contents ranging from 0.95 to 7.43 (average 4.34) ppm. The fibrous to acicular pyrolusite there differs somewhat from that of the usual ore samples in being commonly intergrown with glass-clear barite and the ores are traversed by some calcite and quartz veinlets or encrustations. The variation in radioactivity of the individual ore bodies could be ascribed to several causes none of which can be proved as yet. This variation might be due to an initial local difference in composition of the mineralizing solutions in the different channels or at different times. It could be due to local physico-chemical changes peculiar to each channel such as local contamination

from certain Carboniferous black shales, local variations in temperature, pressure, pH and redox potential, gel-sol conditions, and reactive or adsorptive effects of host rocks encountered by the hydrothermal solutions. Uranium in solution might also have diffused outwardly from some of the Mn-Fe mineralized zones and washed into the adjoining shales, siltstones, and sandstones. These shales and siltstones are liable to display sudden and conspicuous changes in their radioactivities that are not consistent with a normal syngenetic distribution of radioelements. For example, some of the red stained shales above the ore in Om Bogma mines have eU_3O_8 contents as high as 0.077 percent, but a few feet nearby in the same shale this value suddenly falls down to 25 ppm. Some redistribution of the original uranium in these shales and siltstones must also have occurred under the influence of migratory trends of intrastratal solutions set into play by the igneous eruptions and the hydrothermal volcanic activities. Ferruginous (later?) ochreous replacement pockets in the dolomites and limestones are also somewhat radioactive. Maximum radioactivity observed from only four composite samples of such pockets corresponds to an eU_3O_8 content of about 12.5 ppm.

In west central Sinai, there are indications of protracted periods of mineralization or of renewed surges of rather different hydrothermal activities. These are suggested by the alternation of faulting and mineralization, rejuvenation of faults, caverneous habit of some parts of the dolomites and the manganese-iron deposits, new generations (recrystallization?) of coarse pyrolusite traversing pyrolusite ores of different textures in well defined sharp veinlets, by the occasional relics of pyrolusite replaced by hematite, by veins of calcite or silica traversing or lining the cavities of the ores, and finally by the encrustations of turquois (at Maghara) or of malachite on joint surfaces of sandstones, siltstones and shales previously stained by iron or iron-manganese oxides. The chronological order of deposition of copper ores (previously mentioned) and manganese-iron ores of west central Sinai has not been worked out yet but the turquois and malachite encrustations and stains on Fe-Mn stained joints just mentioned might suggest that the copper was transported in a hydrothermal phase later to that depositing the manganese and iron. This is contrary to the tacitly assumed but so far not proved order of mineralization in west central Sinai unless there were renewed rejuvenation of some older copper deposits.

Lead-Zinc Mineralization at Om Gheig.—On the Red Sea coastal strip between El Qoseir and Bir El Ranga (about latitudes 26° and 24° 20′ respectively), lead-zinc mineralization in Miocene sediments occurs in disconnected localities on a linear arrangement parallel to and close to the coast. Mineralization must have taken place in localized spots and fissures along a subsurface long fault or along an important tensional fissure parallel to the Red Sea coast, itself originally the side of a rifted graben. Amin (1, p. 223–225, 236–239) has briefly described these lead-zinc deposits and sketched their stratigraphical relations. He considered them to be of a telethermal replacement origin and he summarily related them to Tertiary volcanic activities (1, p. 225). Some, presumably Oligocene, diabasic flows occur in the coastal strip between latitudes 25° 55′ and 25° 45′ but no direct relationship

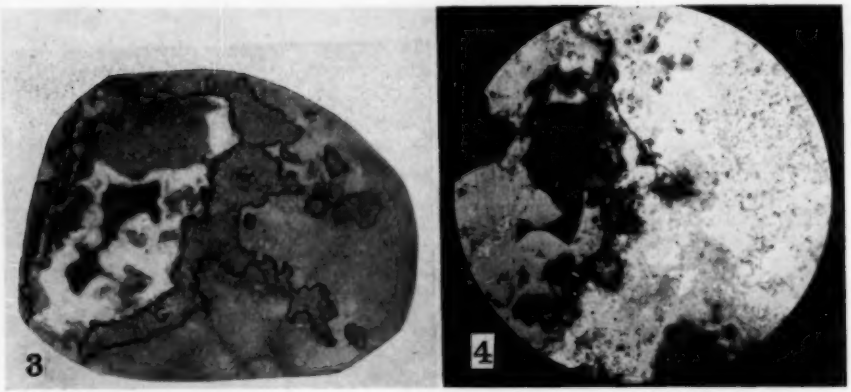

FIG. 3. A polished slab of a Miocene marl (gray) showing irregular replacements by fresh galena (G, boundaries traced with black ink) with some marginal barite. Colloform chalcedony, with an outermost layer of small quartz crystals, lines the dark vug at left. Total radioactivity of a powdered large sample of this rock is 235×10^{-6} alphas/cm^3/sec. Similar samples with hydrothermally altered galena are at least four times more radioactive suggesting a late uraniferous hydrothermal phase of alteration. Om Gheig mines, about twice natural size.

FIG. 4. Photomicrograph of a polished section of a partially altered galena-sphalerite ore from Om Gheig. To the right fairly fresh galena (white) is intergrown with fresh sphalerite (light gray) with their alteration products to the left. The rock sample consists of disconnected small relict shining cores of galena-sphalerite surrounded by wide aureoles of alteration. The average radioactivity of the sulfide cores is 263×10^{-6} alphas/cm^3/sec, while that of their alteration aureoles is $1{,}693 \times 10^{-6}$ alphas/cm^3/sec (range from 610 to $3{,}100 \times 10^{-6}$). Reflected polarized light, $\times$ 7.6.

between them and the lead-zinc mineralizations are available. Rare veinlets of chalcedony and malachite (considered as originally hypogene) were recently discovered lining joint planes in one of these diabases (14, p. 50–51).

The primary sulfide ores are galena, sphalerite intergrown with galena (in places showing signs of replacing it), and of a little pyrite. They occur in different patterns of replacement and/or fissure fillings in the Miocene calcareous gravels, sandstones, marls and clays (e.g. Fig. 3). The mineralized sediments lie below a cap-bed of massive Middle Miocene Gypsum, itself little or non-mineralized. Spectrographic analysis of the galena of Om Gheig deposits suggests that it had been formed at very low temperatures, "specifically below the leptothermal range," (4, p. 38). Rich and diversified varieties of oxidation and alteration products have arisen from the sulfides. These include arsenian wulfenite (4, p. 38), smithsonite, cerussite, hemimorphite, zinc sulfates, limonitic and hematitic ochres and other still unidentified minerals. The deposits at Om Gheig are considered as potentially the most important of the lead-zinc mineralization.

Radioactivity studies on several samples from Om Gheigh indicated that there is *always* a sharp difference in radioactivity between the *fresh* sulfides

(checked by polished sections) and their *alteration products* (Fig. 4). Even powders of thin film alteration products scraped from fracture surfaces inside otherwise fresh lenses of galena have an average alpha radioactivity of 1211 while that of the galena is 228. On the whole the observed radioactivity of the fresh sulfides ranges between 152 and 265 while that of their alteration products progressively increases with alteration to an observed maximum of 2400. Arsenian wulfenite scales (Fig. 5) picked up from some soft porous ochreous altered zones have an average radioactivity of 1671 corresponding roughly to an equivalent U_3O_8 content of about 7.3 ppm. The remaining residue of the sample from which most of the wulfenite scales have been picked up, is still more radioactive than the wulfenite, especially in the coarser size grades (above mesh 70) where the radioactivity rises to 3000. This would indicate the presence of particles (or of a still unidentified mineral) much more radioactive than the wulfenite. Almost pure pockets of ferruginous yellow, red and brown ochres in the Om Gheig mineralized sediments are also radioactive (range observed 514–1070). The host rocks apparently least affected by the different mineralizations and their alteration products have a dominantly low radioactivity. The limegrits for example have observed radioactivities in the range 150–370 increasing to 900 in the limonitic wall rocks.

The presence of arsenian wulfenite and the radioactivity studies indicate that in the mineralizations at Om Gheig there was a late stage surge of

FIG. 5. Orange to red colored tetragonal platy crystals of arsenian wulfenite picked from some altered lead ores at Om Gheig. Transmitted light, × 11.

ascending waters carrying some Mo, As and U and attacking the sulfides deposited earlier. The alteration of the lead-zinc sulfides should not be entirely attributed, as it is until now, to ordinary surficial meteoric oxidation of the primary sulfides.

ECONOMIC URANIUM DEPOSITS IN EGYPT

Apart from indigenous uranium in certain Carboniferous and Cretaceous black shales and lignites and the Maestrichtian phosphates, some of which may eventually prove to be exploitable, Higazy et al. (12) have recently briefly reported four occurrences enriched in uranium and constituting workable ore deposits. The first of these is in the Gebel Qatrani area, (Fig. 1). The general succession of the fluviomarine Oligocene series of Gebel Qatrani, as described by Beadnell (5), consists of about 280 m of alternating variegated sands, sandstones, and calcareous grits intercalated by marls, clays, and lignitic shales, and lies unconformably over the more calcareous Upper Eocene. The lower half of the succession displays estuarine conditions of deposition and yielded a rich fossil content of flora and fauna that include fishes, reptiles (turtles, turtoises, crocodiles, etc.), birds and land mammals (ancestral elephants, etc.). The Oligocene succession is overlain by one of the longest Tertiary basalt sheets and flows, shown on the 1 : 2 million geological map of the country. They extend intermittently to the northeast to Cairo, probably indicating a major subsurface fissure or fault.

Higazy et al. (12, p. 97–98) reported siliceous and uraniferous vertical or steep "veinlets" (joint fillings) in the multicolored Oligocene sandstone near Gebel Qatrani. These "veinlets" contain about 0.3% eU_3O_8, the uranium being concentrated in interstitial spaces between the sand grains. The authors attributed these radioactive "veinlets" to hydrothermal solutions ascending through pre-existing fissures and joints. Black shales and lignitic bands in the succession have an original indigenous radioactivity detectable by ordinary field counters.

The other three uraniferous occurrences mentioned by Higazy et al. (12, p. 98–99) all come from El Qoseir area and near the Red Sea coast. They are apparently restricted to mineralized walls of Miocene faults that had acted as channelways for the uraniferous solutions.

It is suggested here that all these mineralized uraniferous occurrences were primarily caused by hydrothermal solutions connected with the Tertiary volcanic activities of these areas. The acid uprising volcanic solutions could leach and carry upwards the indigenous uranium from the lower stratigraphic horizons of initially low uranium contents (viz. the lower Oligocene fluviomarine shales and lignites of Gebel Qatrani series and the underlying euxenic and carbonaceous Cretaceous black shales and phosphates of the Qoseir area), eventually depositing it in and around favorable channels at higher levels. If conditions permitted, the volcanic solutions could also cause local migrations and concentrations of uranium within the uraniferous mother beds themselves. Such an interpretation does not conflict with the authors' views and is in harmony with the theme of this paper.

SUMMARY AND CONCLUDING REMARKS

1. Diverse hydrothermal volcanic activities followed in the wake of Tertiary simatic magmatic fissure eruptions in Egypt and long survived them. Some of these mineralizing activities produced some pyrite, copper, manganese-iron, and lead-zinc deposits. Although of a relatively very low tenor, the radioactivity of these deposits indicates a close relation with certain phases of the hydrothermal activities.

2. As simatic rocks are the least radioactive of all igneous rocks, vapors and hydrothermal solutions indigenous to such Tertiary magmas must have been originally non-uraniferous or exceedingly poor in uranium. The volcanic solutions must have acquired their uranium during their ascent through the crustal rocks by solution and mobilization of any pre-existing uraniferous deposit or by leaching of any labile uranium in intergranular spaces (or otherwise) from the crystalline, commonly granitic, Precambrian basement and its overlying sedimentary cover. For example, large allanite crystals in the Precambrian pegmatites of Wadi El Gemal are observed to lose readily about two thirds of their radioactivity (largely incurred by loss of their uranium contents) through hydrothermal alteration (personal observation).

In this respect, uranium resembles many of the other mineralizing ions carried by the hydrothermal Tertiary solutions in being of an extraneous origin.

3. Under favorable physico-chemical conditions the transported uranium would be finally deposited. This would naturally depend, among other things, on the intimate properties of the solutions and of the host rocks. In certain cases, economic uraniferous deposits would be formed.

4. The uraniferous deposits would tend to be selectively located at or around planes of discontinuity such as along faults, joints, fissures, unconformities and contrasting contacts, although again the character of the host rocks and the water-table could modify this to a certain extent.

5. The heat of the igneous intrusions could in certain cases "vaporize" and mobilize the pore solutions of adjacent invaded sediments producing hot solutions carrying with them any labile uranium or other leachable ions present. In the Slick Rock district of southwestern Colorado, Shawe et al. (15) concluded that the uraniferous deposits there were initially formed by ground water heated and set into circulation by igneous intrusions.

6. The rather even distribution of uranium within individual samples of malachite and manganese-iron oxide ores of west central Sinai suggests that uranium was coprecipitated from solution with these minerals. This appears also to be the case in the small scale pyrite replacements of chalk and limestone in Kharga Oasis. In the lead-zinc mineralizations of Om Gheig, however, uranium is characteristically associated (precipitated) with the later hydrothermal phases attacking the sulfide minerals deposited earlier.

7. Late supergene meteoric circulations may obliterate or mask the original relation between tectonism, hydrothermal activity and radioactivity by alteration, selective leaching and local redistribution of uranium and other elements.

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ROTATION PROPERTIES OF CERTAIN ANISOTROPIC ORE MINERALS

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ABSTRACT

Rotation properties of 82 anisotropic ore minerals in various wavelengths of light have been measured. Values of the rotation angles for the various minerals range from sensibly 0° to 23.2°. Phase differences (2ϑ) range from negligible to more than 26.3°. The phase differences of most of the minerals are positive. The minerals studied show a variety of patterns of dispersion of both A_r and 2ϑ.

The study indicates that the range of values of rotation properties of ore minerals is sufficiently broad to make these properties useful in identifying minerals with the microscope, particularly when the properties are used in conjunction with other optical properties.

INTRODUCTION

SYSTEMATIC investigation of the rotation properties of anisotropic minerals has been in progress in the laboratory of Economic Geology at the University of Wisconsin since 1955. The present paper presents data for 82 anisotropic minerals as determined by the authors. The data are intended to serve as reference standards for use in the identification of ore minerals.

METHODS OF MEASUREMENT OF ROTATION PROPERTIES

The theory, techniques, and equipment employed in measurement of rotation properties have been discussed by Berek (1), Cameron and Green (5), Cameron, Hutchinson and Green (6), Hallimond (8), and Cameron (3). In general, two quantities are determined for every mineral in each of a series of wavelengths of light. The first is the angle of apparent rotation A_r, which is the angle between the major axis of the reflected ellipse of vibration and the plane of the polarizer when the mineral is at a 45° position, and the second is 2ϑ, the phase difference between the components of the ellipse of vibration at 45° to the ellipse axis. The angle A_r is measured by means of a rotating analyzer, the position of which can be read to 0.1°. The phase difference 2ϑ is calculated from the setting, at compensation, of a rotating mica plate. The setting of the mica plate can likewise be read to 0.1°. The formula used (Fig. 1) is

$$\tan 2\vartheta = \tan \Delta_G \sin 2\alpha_G$$

where Δ_G is the phase difference of the mica plate for the wavelength of light employed.

In the measurement of A_r and 2ϑ (3), the nicols are crossed, and the extinction positions of the mineral grain under observation are found. In general, there will be four extinction positions 90° apart if the nicols are exactly crossed. At extinction, the vibration directions of the grain are parallel to the vibration planes of the nicols. The stage is now turned 45°,

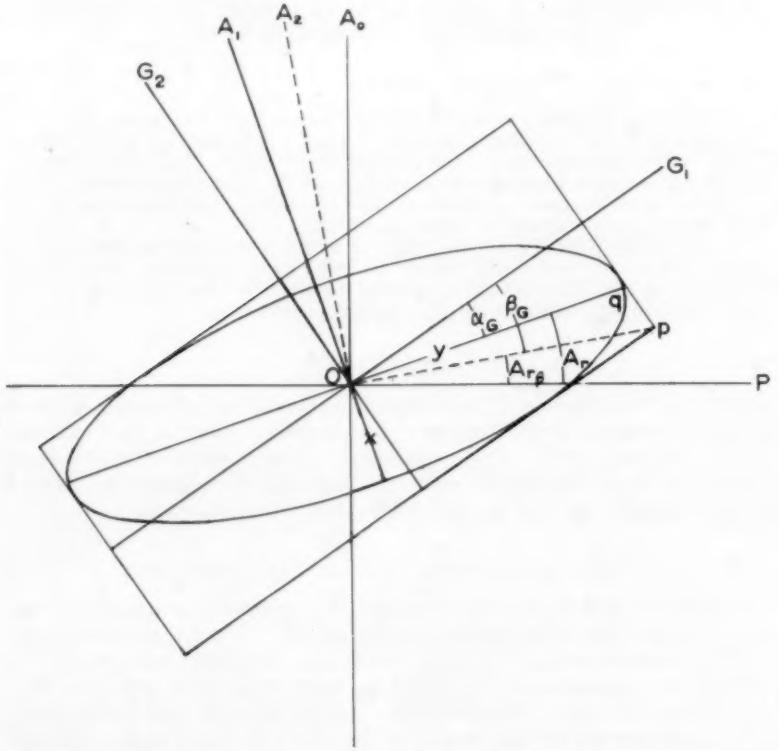

FIG. 1. Diagram showing relations of analyzer and mica compensator to a reflected ellipse of vibration, at compensation. OP—vibration plane of polarizer; OA_0—vibration plane of analyzer at crossed position; x, y—semi-minor and semi-major axes of ellipse; $y = Oq$; $A_1 \perp Oq$, angle $A_1OA_0 = A_r$; $A_2 \perp Op$, $A_2OA_0 = A_{r\beta}$; OG_1—slow ray direction of mica plate; OG_2—fast ray direction of mica plate.

either clockwise or counterclockwise, so that the mineral vibration directions are at 45° to the nicol vibration planes. A Nakamura plate is inserted into the optical system (3, 8) and the analyzer is turned until the two halves of the Nakamura plate are equally dark. The vibration plane of the analyzer (A_1, Fig. 1) is now perpendicular to the major axis of the reflected ellipse of vibration. The angular rotation of the analyzer (A_0OA_1) is the apparent

angle of rotation A_r. The analyzer is left at the A_1 position, the ocular and Nakamura plate are removed, a pinhole ocular is inserted, and a 6-mm or 4-mm objective is used. The polarization figure is now seen. The rotating mica plate is now inserted, and it and the analyzer are turned until the isogyres of the figure form a black cross. The setting of the mica plate is now read, and the angle through which it has been turned from its zero position calculated. From A_r and the setting of the mica plate the angle α_G is calculated (Fig. 1). Three separate determinations of A_r and the corresponding values of α_G are made, and the average value of each is calculated. The grain is now turned 90° by means of the stage, to a second 45° position. The measurements are repeated. The means of the average values of A_r and α_G at the two 45° positions are taken as the final values.

The sensitivity of the Nakamura plate is such that A_r for most minerals can be measured with remarkable accuracy. For values up to 5°, A_r can almost always be measured with an accuracy (mean deviation) of ±0.1° to ±0.2°, and sometimes to ±0.05°. For values above 5°, accuracy for some minerals is less, ±0.3° to ±0.4°. For angles of 15° or higher, errors are 0.5° or even larger for a few minerals that produce large phase differences. For umangite and rickardite, A_r cannot be measured, owing to high phase differences, and the Berek method must be employed (3). By this method, the angle $A_{r\beta}$ (Fig. 1) and the angle β_G are measured. The angle β_G is used to calculate 2ϑ by the formula:

$$\sin 2\vartheta = \sin 2\beta_G \sin \Delta_G$$

where Δ_G is the phase difference of the mica plate for the wavelength of light employed. Then

$$\cos 2\alpha_G = \frac{\cos 2\beta_G}{\cos 2\vartheta}$$

and, for the case shown in Figure 1,

$$A_r = \sphericalangle G_1OP - \alpha_G = \sphericalangle G_2OA_0 - \alpha_G.$$

The reader is referred to the paper by Cameron (3) for further details of the calculations.

All measurements presented in this paper have been made in illumination furnished by a carbon arc lamp. Monochromatic light of various wavelengths is produced by means of Bausch and Lomb second order interference filters (half width 8–10 mμ), supplemented where necessary by filters to eliminate secondary peaks. Auxiliary filters used are Corning Nos. 3-69, 3-73, 4-97, and Pittsburgh No. 2043. Measurement in a given wavelength is the more accurate the greater the intensity of illumination. With the carbon arc lamp, intensities are entirely satisfactory within the range 500 to 620 mμ. For minerals of low reflectivity, intensity is low below 500 mμ and above 620 mμ, hence measurements for some minerals at 470 and 650 mμ are of lower accuracy. At 620 and 650 mμ, certain minerals show internal reflections that

prevent measurement of A_r and 2ϑ. At 520 to 600 mμ, satisfactory measurements can be made for most ore minerals with only a low voltage tungsten lamp as the source.

Values for 2ϑ are a function of precision in setting the mica plate at the position of compensation. This can sometimes be done to ±0.1° to 0.2°. With some minerals, however, precise setting is impossible with present equipment, and in some cases the errors in setting are as much as ±0.5 to 1.0°. Depending on wavelength these correspond to errors of 1.1 to 4.8° in the calculated value of 2ϑ. Even where 2ϑ cannot be determined with reasonable accuracy, however, the sign of the phase difference can in most cases be determined and has value in identification. For many minerals, the sign in white light can be determined qualitatively, and very quickly, by means of a full-wave gypsum plate (sensitive tint). The procedure is as follows:

1. With nicols crossed, determine the extinction positions of the grain under observation.
2. Rotate the stage 45°. The mineral is now illuminated.
3. Rotate the analyzer counterclockwise. If the grain becomes brighter, rotate the stage 90° to the next 45° position. The mineral will now become darker as the analyzer is rotated counterclockwise from the crossed position.
4. Restore the analyzer to the crossed position. Insert the gypsum plate. If the color changes to bluish, the sign of the phase difference is positive, if to a yellowish tint, the sign is negative.

In the above, it is assumed that the vibration plane of the polarizer is E–W and the gypsum plate is inserted so that its fast-ray direction is NW–SE. If the polarizer vibration plane is N–S, then in the above a bluish tint indicates a negative sign, a yellowish tint a positive sign.

Phase differences produced by many ore minerals are small, and the changes in the color of the gypsum plate produced are much weaker in general than with the transparent anisotropic minerals in transmitted light. The change from faintly bluish to faintly yellowish can sometimes be seen only if the mineral is rotated from one 45° position to the next, when the contrast becomes apparent. Even so, for certain ore minerals the effect is too faint to be satisfactory.

CORRECTION OF VALUES OF A_r AND 2ϑ

When an anisotropic mineral rotates an incident linear vibration, the reflected vibration undergoes a further rotation in passing upward through the reflecting plate (3; 8, p. 127). A correction of the observed value of A_r for this rotation is always necessary to obtain the true value of A_r. The values given in Table I, except for those in the last column, are corrected values. In actual practice, however, there is a definite advantage in using observed values rather than the corrected values, because the effect of the reflecting plate is to spread the observed values over a broader range than the true values. This

aids identification, particularly since the accuracy of measurement is generally $\pm 0.1°$ to $0.2°$ for the *observed* values. For purposes of comparison, observed values of A_r at 589 mμ are given in the last column of Table I.

Anyone making use of rotation properties for routine mineral identification will find it worthwhile to convert the corrected values given in Table I to observed values appropriate to the microscope he is using and to work entirely in terms of observed values. Once such a table of observed values has been drawn up, calculation of true from observed values is no longer necessary.

For maximum accuracy, correction curves must be determined for a given microscope for each of the wavelengths used. The variation in the correction with wavelength is small for the three Bausch and Lomb microscopes used by the authors in the present work, only 0.8° over the range of 470 mμ to 650 mμ even for observed angles of 25°. For angles below 5° the variation with wavelength is negligible. The ratio of observed to corrected angles for all wavelengths is close to 6:5, the ratio reported by Hallimond (8, p. 107) for a Cooke ore microscope.

The quantity 2ϑ is calculated from the setting of the mica plate (Fig. 1). The angle α_a must be corrected by the same factor used in correcting A_r. In Table I this correction is incorporated in the value of 2ϑ given. In this connection, it should be noted that for the present the value of 2ϑ should be taken solely as indicating the sense and order of magnitude of 2ϑ, for two reasons. One reason is that the effects of the surface film on values of 2ϑ for various minerals (see below) have not yet been adequately appraised. Berek (1, p. 102) and Turner, Benford, and McLean (10) have indicated that the effects may be serious for some minerals. The second reason is that a small error in the setting of the mica plate means a much larger error in the calculated value of 2ϑ. The error increases in more than direct proportion as the phase difference increases.

Notes on individual minerals of Table I

Arsenopyrite—the maximum value of A_r for sodium light is given by either the *ac* or the *ab* section, but the *bc* section gives maximum values for wavelengths toward the ends of the spectrum, and maximum anisotropism in white light.

Anatase—values measured from prism face of crystal.

Ramsdellite—material from Lake Valley, New Mexico, proved impossible to polish owing to perfect splintery cleavage. Values given were measured from a cleavage face and are subject to revision when polished material is available.

Tapiolite—specimens from different localities give different values. Values given are for tapiolite from the Best Bet pegmatite, Yellowknife district, N. W. T., described by R. W. Hutchinson (8).

Graphite—A_r not measurable with accuracy, apparently owing to smearing of polished surfaces of graphite crystals with minute flakes of graphite. Value given for Na light is probably lower than the true maximum.

Cassiterite—values measured on prism face of crystal.

Hausmannite—maximum values. A_r varies with polishing.

Covellite—values of A_r in a normally polished surface. With careful buffing, values are increased (see text).

Pearceite—A_r varies irregularly with repeated polishing.

Polybasite—A_r varies irregularly with repeated polishing.

TABLE I - ROTATION PROPERTIES OF ORE MINERALS [a]

Mineral	Obs.	470 nm		520 nm		546 nm		589 nm		620 nm		650 nm		589 nm Gil	[b] Reflectivity in Percent	Effect Of Gypsum Plate [c]	Observed Δ_T 589 nm
		Δ_T	2δ	Δ_T	2δ	Δ_T	2δ	Δ_T	2δ	Δ_T	2δ	Δ_T	2δ	Δ_T			
Platnerite	B														17.0		0.0
Arsenopyrite (bx section)	L	1.33 ± .1	-2.4 ± .9	0.96 ± .1	-3.0 ± .3	0.73 ± .1	-2.1 ± .6	0.0 ± .1	-1.2 ± .1	-0.17 ± .1	+2.9 ± .2	-0.25 ± .1	+1.3 ± .4	- - -	52.4	(-)D	0.0 ± .1
								0.2						- - -	50	E	0.2
Nuncherite	C	---	---	---	---	---	---	---	---	---	---	---	---	---	60.0	E	0.2
Aguilarite	L	0.39 ± .1	-3.6 ± .4	0.24 ± .1	-2.7 ± .3	0.16 ± .1	-2.1 ± .6	0.16 ± .1	-1.8 ± 1.3	0.17 ± .1	-1.2 ± .2	0.08 ± .1	nil	0.32 ± .1	29.9	--	0.2 ± .1
								0.32 ± .1	---	---	---	---	---	- - -	238.0	E	0.4 ± .3
Stromeite	D	---	---	---	---	---	---	---	---	---	---	---	---	---			
Chalcopyrite	C	---	---	---	---	---	---	0.32 ± .1	---	---	---	---	---	---	43.2	E	0.4 ± .2
Zincite	D	---	---	---	---	---	---	0.32 ± .1	---	---	---	---	---	---	21.0	E	0.4 ± .2
Wagnerite	D	0.98 ± .1	+1.6 ± 1.3	0.72 ± .1	+1.6 ± 1.4	0.67 ± .1	+1.6 ± .9	0.57 ± .1	1.1 ± .3	0.33 ± .1	nil	0.25 ± .1	nil	---	33.2	E	0.7 ± .1
Brannite	S	0.70 ± .2	nil	0.72 ± .1	nil	0.73 ± .1	nil	0.65 ± .1	nil	0.66 ± .1	nil	0.67 ± .1	+0.4 ± .2	1.06 ± .1	80.1	(+)F	0.8 ± .1
Antimony	L	0.63 ± .1	+5.2 ± .4	0.72 ± .1	+4.5 ± .3	0.81 ± .1	+4.2 ± .3	0.81 ± .1	0.8 ± .4	0.99 ± .1	+3.5 ± .2	1.17 ± .1	+3.4 ± .4	1.17 ± .1	74.4	(+)F	1.0 ± .1
Bismuthinite	O	2.88 ± .3	nil	2.08 ± .1	-4.8 ± .9	---	---	0.81 ± .1	-1.9 ± .4	0.58 ± .1	-2.5 ± .5	0.33 ± .1	---	---	52.6	(-)F	1.0 ± .1
Stannite	M	1.56 ± .1	-2.1 ± 1.3	0.96 ± .1	-1.1 ± .7	0.89 ± .1	nil	0.89 ± .1	nil	0.82 ± .1	-0.7 ± .6	0.50 ± .1	-2.1 ± .4	2.44 ± .1	27.6	(-)F	1.1 ± .1
> varamaleburgite	M	0.68 ± .1	+3.6 ± .9	0.32 + .1	+5.1 ± .7	0.48 ± .1	+3.7 ± .3	0.89 ± .1	+4.9 ± .6	1.16 ± .1	+3.7 ± .5	1.25 ± .1	+2.8 ± .4	1.30 ± .1	-	(+)D	1.1 ± .1
. : tee (crystal)	M	0.94 ± .1	+0.9 ± 1.2	0.96 ± .1	+1.4 ± .7	0.89 ± .1	nil	0.89 ± .1	+0.1 ± .2	0.82 ± .1	+0.5 ± .5	0.91 ± .1	+1.0 ± .6	---	-	E	1.1 ± .1
Naiadite	M	-0.19 ± .8	-18.5 ± 1.3	0.80 ± .2	+11.5 ± .7	0.81 ± .2	+9.0 ± .6	0.89 ± .1	+7.1 ± .6	0.82 ± .1	+6.1 ± .7	0.91 ± .2	+5.8 ± .4	1.30 ± .2	-	(+)D	1.1 ± .1
Arsenic	O	0.35 ± .1	+1.3 ± .9	0.80 ± .1	+1.4 ± .7	0.97 ± .2	+1.4 ± .3	0.97 ± .2	+1.1 ± .2	0.82 ± .2	+0.9 ± .2	---	---	1.72 ± .2	52.6	(+)D	1.2 ± .2
Arsenopyrite	L	0.86 ± .1	+1.2 ± .9	0.80 ± .1	+3.0 ± .5	0.73 ± .1	+3.0 ± .3	0.97 ± .1	+2.7 ± .4	1.16 ± .1	+2.3 ± .5	1.33 ± .1	+1.9 ± .4	1.63 ± .1	36.7	(+)D	1.2 ± .2
Polybasite (ordered)	L	0.94 ± .1	+4.8 ± .9	1.44 ± .1	+3.0 ± .7	1.09 ± .1	+1.4 ± .3	0.97 ± .1	+0.2 ± .2	0.82 ± .1	+0.5 ± .5	0.83 ± .1	+0.8 ± .4	1.95 ± .2	34.9	(+)D	1.2 ± .2
Bronzite	O	1.48 ± .2	nil	1.14 ± .1	-5.0 ± 1.4	1.05 ± .1	---	0.97 ± .7	-0.8 ± .4	0.91 ± .1	-0.1 ± .5	0.83 ± .1	---	2.92 ± .5	15.6	(-)F	1.2 ± .2
Tantalite	O	0.86 ± .1	+4.0 ± .2	0.96 ± .1	+1.4 ± .7	1.05 ± .1	+1.1 ± .6	1.06 ± .2	nil	0.99 ± .1	+0.7 ± .7	0.83 ± .1	nil	2.88 ± .5	18.5	(+)F	1.3 ± .2
Jezeite	M	0.86 ± .1	+6.0 ± 1.3	1.08 ± .1	+4.5 ± .3	1.05 ± .1	+3.5 ± .6	1.14 ± .1	+3.3 ± .2	1.32 ± .1	+2.5 ± .5	1.42 ± .2	+2.2 ± .4	1.38 ± .1	-	(+)D	1.4 ± .1
Safflorite	M	0.70 ± .1	-15.5 ± 1.3	0.84 ± .1	+12.5 ± .7	0.81 ± .2	-11.5 ± .6	1.14 ± .1	+7.4 ± .5	0.66 ± .2	---	1.33 ± .1	+5.4 ± .4	2.44 ± .2	58.0	(+)D	1.4 ± .1
Tetradymite	O,D	0.88 ± .1	+0.4 ± .8	0.96 ± .1	+2.6 ± .3	1.13 ± .1	+1.4 ± .6	1.14 ± .1	+1.4 ± .5	1.26 ± .1	+0.2 ± .2	1.33 ± .1	+0.4 ± .4	1.79 ± .2	56.9	(+)D	1.4 ± .1
Pyrargyrite	L	0.63 ± .1	+1.6 ± .9	0.96 ± .1	+1.4 ± .3	0.97 ± .1	+1.1 ± .6	1.14 ± .1	+1.8 ± .5	0.97 ± .1	+1.2 ± .9	0.83 ± .2	---	1.63 ± .2	30.8	(+)D	1.4 ± .1
Geocronite	M	1.33 ± .2	+ small	1.36 ± .1	+ small	1.29 ± .2	+0.6 ± .6	1.22 ± .1	+ small	1.16 ± .1	+ small	1.08 ± .1	+ small	1.99 ± .2	40.9	(+)D	1.5 ± .1
Orlandite	O	1.17 ± .1	+1.2 ± .4	1.20 ± .1	nil	1.21 ± .1	nil	1.30 ± .1	nil	1.24 ± .1	nil	1.25 ± .1	nil	2.92 ± .1	18.5	(+)F	1.6 ± .1
Hinginite	L	1.25 ± .1	+1.6 ± .4	1.36 ± .1	+1.6 ± .3	1.45 ± .1	+1.1 ± .3	1.46 ± .1	+0.9 ± .3	1.48 ± .1	+0.7 ± .5	1.33 ± .1	+0.6 ± .4	2.36 ± .3	44.0	(+)D	1.8 ± .1
Stephanite	L	-0.39 ± .1	-3.6 ± .9	-0.16 ± .1	+5.2 ± .7	0.65 ± .1	+4.8 ± .6	1.46 ± .1	+3.8 ± .5	1.65 ± .1	+1.9 ± .2	1.50 ± .1	+0.8 ± .4	2.28 ± .2	29.7	(+)D	1.8 ± .1
Cylindrite	M	0.94 ± .2	+1.3 ± 1.3	1.14 ± .2	nil	1.25 ± .1	nil	1.46 ± .1	+0.9 ± .5	1.65 ± .2	+0.9 ± .7	1.75 ± .1	+0.4 ± .4	2.36 ± .3	33.0	(+)D	1.8 ± .1
Cassiterite	O	1.33 ± .1	+0.9 ± .9	1.44 ± .1	+1.1 ± .7	1.49 ± .1	+1.4 ± .6	1.46 ± .1		1.49 ± .1	+0.5 ± .5	1.42 ± .1	---	3.66 ± .1	12.9	(+)D	1.8 ± .1
Pyrrhotite	O	1.95 ± .1	-1.6 ± .9	1.84 ± .1	-2.1 ± .3	1.69 ± .1	-2.8 ± .6	1.55 ± .5	-3.2 ± .5	1.24 ± .1	-1.7 ± .2	1.25 ± .1	-3.2 ± .6	2.13 ± .1	39.9	(-)D	1.9 ± .1
Franckite	L	1.01 ± .1	+2.1 ± 1.3	1.44 ± .1	+1.6 ± .7	1.45 ± .1	+1.4 ± .6	1.65 ± .1	+1.0 ± .3	1.65 ± .1	+1.1 ± .5	1.58 ± .2	+0.4 ± .4	2.60 ± .2	29.9	(+)D	2.0 ± .1
Sinkenite	D	1.82 ± .2	+2.8 ± 1.3	1.60 ± .2	+1.8 ± .7	1.60 ± .2	+1.1 ± .6	1.65 ± .1	+0.8 ± .5	1.69 ± .1	+1.1 ± .5	1.91 ± .2	+1.5 ± .4	2.84 ± .3	38.0	(+)D	2.0 ± .1
Haromeite	L	2.36 ± .1	nil	2.44 ± .1	-2.7 ± .3	2.09 ± .1	-3.7 ± .3	1.65 ± .1	-3.5 ± .5	1.49 ± .1	-3.1 ± .2	1.33 ± .1	-2.6 ± .2	---	51.5	(-)D	2.0 ± .1
Psaronite	L	0.78 ± .1	+0.9 ± .9	1.44 ± .2	+2.7 ± .7	1.62 ± .1	+4.2 ± .6	1.65 ± .2	+3.6 ± .5	0.82 ± .2	+2.3 ± .5	0.58 ± .1	+1.2 ± .4	---	30.2	(+)D	2.0 ± .2
Polybasite (disordered)	L	0.15 ± .1	+5.6 ± .4	1.08 ± .1	+4.8 ± .7	1.25 ± .2	+3.2 ± .3	1.71 ± .2	+2.1 ± .5	1.82 ± .2	+1.1 ± .2	1.50 ± .2	+0.4 ± .2	2.76 ± .2	-	(+)D	2.1 ± .2
Famatinite	L	1.01 ± .1	nil	1.28 ± .1	nil	1.45 ± .1	nil	1.79 ± .1	nil	2.15 ± .1	-0.9 ± .5	1.83 ± .2	-0.6 ± .4	3.48 ± .1	26.2	E	2.2 ± .1
Millerite	O	0.0 ± .2	+14.0 ± .9	1.20 ± .1	+11.0 ± 1.1	---	---	1.79 ± .1	+4.1 ± .5	1.74 ± .1	+6.4 ± .7	---	---	2.52 ± .2	59.5	(+)D	2.2 ± .1
Rathcetite	L,S	1.67 ± .1	+1.5 ± .4	2.00 ± .1	+1.4 ± .1	2.02 ± .1	+1.3 ± .3	2.03 ± .1	+1.1 ± .3	2.07 ± .1	+1.0 ± .2	2.00 ± .1	+1.1 ± .4	---	17.1	(+)D	2.5 ± .1
Aikinite	L	1.54 ± .1	+3.2 ± .4	1.76 ± .1	+2.4 ± .3	1.93 ± .1	+2.1 ± .6	2.03 ± .1	+1.6 ± .5	1.98 ± .1	+1.1 ± .5	1.91 ± .1	+0.2 ± .2	3.09 ± .1	36.0	(+)D	2.5 ± .1

Bismuth	L	2.03 ± .1	+5.2 ± 1.3	2.16 ± .2	+3.0 ± .7	2.38 ± .2	+2.9 ± .6	2.12 ± .2	+1.2 ± .3	1.82 ± .2	+0.9 ± .6	1.67 ± .1	+0.8 ± .6	—	71.3	(+)B	2.6 ± .2
Bismarckite	L	1.01 ± .1	+0.9 ± .4	1.90 ± .2	+1.4 ± .3	1.85 ± .1	+0.9 ± .3	2.12 ± .1	-0.7 ± .3	2.15 ± .1	-1.2 ± .5	1.83 ± .2	-1.5 ± .4	3.42 ± .2	28.0	(-)B	2.6 ± .1
Bismutite	L	1.82 ± .1	+3.2 ± .9	2.00 ± .1	+2.1 ± .3	2.09 ± .1	+1.8 ± .6	2.12 ± .1	+0.9 ± .3	2.07 ± .1	+0.9 ± .5	2.00 ± .2	nil	3.42 ± .1	35.9	(+)B	2.6 ± .1
Preiselebenit	L	1.64 ± .1	-1.3 ± 1.3	1.92 ± .1	-1.4 ± 1.1	2.09 ± .1	nil	2.19 ± .1	+0.7 ± .5	2.32 ± .1	+1.4 ± .2	2.42 ± .1	+2.2 ± .3	4.22 ± .2	17.9	(+)D	2.7 ± .1
Perberite	M	1.64 ± .1	nil	2.08 ± .1	+3.0 ± .7	—	—	2.19 ± .2	+0.5 ± .5	2.07 ± .1	+0.7 ± .7	—	—	3.73 ± .2	27.2	(+)D	2.7 ± .2
Hematite	G	2.26 ± .1	nil														
Vegielite	M	1.17 ± .1	+ small	1.84 ± .1	+ small	2.08 ± .1	+ small	2.28 ± .1	+ small	2.32 ± .1	+ small	2.17 ± .1	+ small	4.06 ± .2	—	(+)B	2.8 ± .1
Tellurium	L	2.50 ± .1	-4.4 ± .1	2.40 ± .1	-2.7 ± .7	2.26 ± .2	-2.8 ± .3	2.28 ± .1	-1.8 ± .3	2.39 ± .1	-1.7 ± .5	2.42 ± .2	-0.6 ± .2	3.25 ± .2	56.5	(-)P	2.8 ± .1
Batile	G	1.87 ± .2	nil	2.16 ± .1	+0.7 ± .7	—	—	2.28 ± .2	+0.9 ± .5	2.07 ± .1	-0.3 ± .9	—	—	3.25 ± .5	22.9	(+)B	2.8 ± .2
Ilmenite	G	2.54 ± .3	nil	—	—	2.58 ± .1	-0.6 ± .3	2.36 ± .1	-1.2 ± .5	2.07 ± .1	-0.5 ± .1	—	—	3.98 ± .2	19.4	(+)B	2.9 ± .1
Meteorolite	G	—	—	2.40 ± .1	+1.6 ± .6	2.42 ± .3	+1.4 ± .9	2.24 ± .1	+0.7 ± .7	—	—	2.42 ± .2	nil	4.71 ± .3	13 ± 17.7		3.0 ± .1
Quenselite	G	2.73 ± .2	+6.0 ± .4	2.64 ± .1	+1.6 ± .4	2.50 ± .2	+1.4 ± .6	2.52 ± .3	+2.5 ± .5	2.56 ± .2	+1.2 ± .9	2.56 ± .2	+2.8 ± .4	5.78 ± .3	18 - 22	(+)D	3.1 ± .3
Cinnaber	D	2.54 ± .1	+8.0 ± 1.3	2.96 ± .1	+2.1 ± .7	2.98 ± .1	+1.8 ± .6	2.52 ± .1	+1.2 ± .5	—	—	—	—	—	26.3	(+)D	3.1 ± .1
Boulangerite	M	2.54 ± .1	+2.1 ± .9	2.80 ± .1	nil	2.62 ± .1	+0.6 ± .6	2.52 ± .2	+1.2 ± .5	2.39 ± .3	nil	2.08 ± .2	-1.3 ± .4	3.89 ± .1	39.4	(+)D	3.3 ± .2
Jamesonite	M	2.03 ± .2	+6.0 ± .9	2.24 ± .1	+4.5 ± .7	2.34 ± .1	+3.8 ± .6	2.60 ± .1	+2.7 ± .5	2.56 ± .1	+2.3 ± .5	2.42 ± .1	+1.7 ± .6	4.22 ± .2	37.7	(+)D	3.2 ± .1
Cervantite	G	3.04 ± .3	+5.2 ± 1.3	—	—	2.51 ± .2	+0.9 ± .3	2.76 ± .2	+2.2 ± .7	2.81 ± .2	—	2.50 ± .2	nil	4.31 ± .3	30.6-34.4	(-)D	3.4 ± .2
Aikinite	L	2.18 ± .1	+2.4 ± .9	2.56 ± .1	+1.6 ± .3	2.74 ± .1	+1.6 ± .3	2.84 ± .1	+0.7 ± .3	2.81 ± .1	+0.5 ± .2	2.67 ± .1	—	4.56 ± .4	45.8	(+)D	3.5 ± .1
Bolafossite	M	2.81 ± .3	+ small	3.20 ± .3	+ small	3.14 ± .2	+ small	2.92 ± .2	+ small	2.81 ± .2	+ small	2.58 ± .3	+ small	5.42 ± .2	22.2	(+)F	3.6 ± .2
Breithauptite	G	2.66 ± .3	+3.2 ± 1.3	4.00 ± .2	+0.7 ± .3	—	—	3.01 ± .1	-1.2 ± .7	2.15 ± .2	-1.1 ± .2	1.67 ± .1	—	—	47.7	(-)F	3.7 ± .2
Sylvanite	M	1.75 ± .1	-2.8 ± .9	3.20 ± .1	-3.6 ± .7	3.31 ± .7	-3.5 ± .8	3.09 ± .1	-3.7 ± .5	2.83 ± .1	-5.1 ± .5	2.75 ± .1	-5.3 ± .4	—	56.3	(-)B	3.8 ± .1
Miargyrite	L	2.18 ± .1	+6.0 ± .9	2.56 ± .1	+4.8 ± .3	2.66 ± .7	+4.0 ± .6	3.09 ± .1	+3.6 ± .3	3.23 ± .2	+2.5 ± .5	3.17 ± .2	+1.7 ± .4	5.41 ± .2	54.0	(+)D	3.8 ± .1
Hollandite	G	3.43 ± .2	-2.8 ± 1.6	3.52 ± .1	-1.8 ± .3	3.48 ± .2	-2.1 ± .3	3.33 ± .2	-1.8 ± .3	—	—	3.17 ± .1	-1.9 ± .2	5.26 ± .2	27.3	(-)D	4.1 ± .2
Cryptomelane	G	3.82 ± .1	-2.1 ± .9	—	—	3.63 ± .1	-1.4 ± .3	3.66 ± .1	-1.6 ± .5	3.55 ± .1	-2.1 ± .5	3.42 ± .1	-2.1 ± .6	5.73 ± .1	29.4	(-)D	4.3 ± .1
Todorokite	G,L	4.21 ± .2	nil	4.16 ± .2	-1.6 ± 1.1	4.03 ± .2	-1.6 ± .8	3.66 ± .1	-1.2 ± .7	3.31 ± .2	-1.1 ± .9	2.81 ± .2	—	5.67 ± .4	—	(+)D	4.5 ± .1
Livingstonite	L	4.13 ± .1	+6.0 ± .6	4.00 ± .1	+3.6 ± .7	3.86 ± .1	+2.1 ± .6	3.81 ± .1	+0.9 ± .5	3.55 ± .1	+0.5 ± .6	3.50 ± .1	nil	6.08 ± .1	31.6	(+)D	4.7 ± .1
Bismuthinite	L	3.28 ± .2	+3.2 ± .4	3.60 ± .1	+2.1 ± .7	3.95 ± .6	nil	3.90 ± .1	nil	3.63 ± .1	-0.5 ± .5	3.42 ± .2	-0.8 ± .6	5.98 ± .2	41.0	(+)D	4.8 ± .1
Pyrolusite	G	3.82 ± .1	nil	3.84 ± .2	+1.1 ± 1.4	3.79 ± .2	-0.9 ± .9	3.98 ± .2	nil	4.13 ± .1	nil	4.25 ± .1	+ small	6.55 ± .1	33.1-39.7	(+)D	4.9 ± .2
Groutite	G	3.90 ± .2	+3.6 ± .8	4.00 ± .1	+3.9 ± .3	4.03 ± .1	+3.2 ± .3	4.06 ± .1	+2.2 ± .3	4.13 ± .1	+1.7 ± .6	3.83 ± .2	+1.5 ± .6	7.48 ± .2	-	(+)D	5.0 ± .1
Manganite	G	4.05 ± .3	nil	—	—	4.03 ± .3	+1.6 ± .6	4.14 ± .2	+ small	4.05 ± .2	+0.7 ± .5	—	—	6.56 ± .3	35.0	(+)D	5.1 ± .2
Manta MnO_2	M	5.82 ± .1	+3.2 ± 1.6	5.12 ± .08	-1.6 ± .7	4.94 ± .1	-1.1 ± 1.1	4.59 ± .1	-2.2 ± .7	4.08 ± .1	-1.4 ± .5	3.63 ± .1	-2.1 ± .2	6.92 ± .2	54.3	(-)D	5.6 ± .1
Berthierite	M,L	5.26 ± .1	+ small	5.19 ± .2	+ small	5.16 ± .2	+ small	4.98 ± .2	+ small	4.81 ± .2	+ small	4.31 ± .2	+ small	6.39 ± .5	19.6	(+)D	5.1 ± .2
Stibnite	L	5.88 ± .2	+12.0 ± .2	5.51 ± .1	+4.8 ± .7	5.41 ± .1	+2.8 ± .6	5.07 ± .2	+2.2 ± .3	4.72 ± .1	+0.9 ± .5	4.57 ± .1	+0.4 ± .2	—	38.9	(+)B	6.2 ± .1
Manganite	B	5.41 ± .2	+ small	5.36 ± .2	- small	5.32 ± .2	- small	5.47 ± .2	- small	5.23 ± .1	- small	5.17 ± .3	- small	10.09 ± .2	19.0-31.4	(-)B	6.7 ± .2
	B	—	—	—	—	8.80 ± .2	-4.5 ± .9	7.35 ± .2	-2.2 ± .3	6.96 ± .3	-2.1 ± 1.5	6.32 ± .2	-3.6 ± .4	10.60 ± .5	-	(-)B	9.0 ± .2
Rosedellite (cleavage)	N	3.43 ± .2	-0.4 ± 1.2	4.72 ± .2	- small	5.82 ± .2	- small	8.22 ± .3	-1.4 ± 1.2	10.42 ± .3	-8.8 ± 1.4	11.85 ± .2	-9.0 ± 1.0	—	-	(-)B	10.1 ± .3
Ilmenite	L	11.12 - .2	+2.4 ± 1.2	10.22 ± .2	+3.6 ± .3	10.85 ± .1	+3.7 ± .6	9.95 ± .1	+4.3 ± .6	10.60 ± .1	+4.9 ± .5	10.62 ± .2	+4.5 ± .4	15.75 ± .3	16-36	(+)B	12.2 ± .1
Molybdenite (oriented)																	
Lithiophorite	G	11.90 ± .6	+11.5 ± 2.4	12.31 ± .5	+9.5 ± 1.6	12.30 ± .4	+6.0 ± 1.4	11.75 ± .4	+4.3 ± .9	11.08 ± .7	+5.1 ± 1.4	11.21 ± .8	+4.5 ± 1.3	—	10.4-19.4	(+)B	14.4 ± .4
Chalcophanite	G	16.41 ± .3	+15.5 ± 3.3	15.10 ± .2	+13.5 ± 4.4	15.20 ± .3	+9.7 ± 1.6	14.52 ± .4	+5.2 ± 1.6	14.1 ± .1	n.d.	13.80 ± .2	+1.3 ± .6	22.9 ± .8	9-30	(+)B	17.8 ± .4
Elaeolite	M	12.13 ± .3	+10.8 ± 1.2	13.35 ± .6	-6.8 ± 1.2	14.25 ± .7	-9.5 ± .7	15.48 ± .6	-1.5 ± 1.1	17.6 ± .20	-12.3 ± 2.0	18.40 ± 1.0	-16.5 ± 2.5	—	19-31	(-)B	19.0 ± .6
Orvallite	L	5.72 ± .2	-6.4 ± 1.2	7.75 ± .2	-3.9 ± .7	7.02 ± .2	-4.6 ± .9	11.91 ± .4	-5.6 ± .5	20.7 ± .4	-7.4 ± .3	23.2 ± .9	+15.0 ± .9	—	9.5-15.1	(-)B	24.6 ± .4
Graphite	G	—	—	—	—	—	—	16.3 ± 80	—	—	—	—	—	—	-		20.0 ± 2.0
Uranglite *	M	1.7 ± .2	-2.3 ± .6	3.2 ± .3	-5.0 ± .8	3.5 ± .3	-6.8 ± .8	0.1 ± .1	-17.4 ± .6	-2.7 ± .2	-80.1 ± .6	-4.8 ± .4	-17.9 ± 1.5	—	15.3	(-)B	0.1 ± .1
Richardsite *	M	6.3 ± .3	-4.6 ± .4	7.9 ± .3	-7.6 ± .8	8.7 ± .2	-18.7 ± .3	4.8 ± .2	-27.1 ± 1.5	-2.7 ± .4	-26.3 ± .8	—	—	—	26-37	(-)B	4.8 ± .2

‡ Values of A_λ and $2\mathcal{V}$ are given in degrees.
* Calculated from values obtained by Berek method.
Obs. = observer: C = Cameron, D = Davis, G = Guilbert, L = Larsen, M = Mansmo, S = Soren.
1 At 550 mμ, for rosedellite A_λ = 10.0 ± .3, $2\mathcal{V}$ = 2.8 ± 2.3; at 500 mμ, A_λ = 9.0 ± .2, $2\mathcal{V}$ = -8.5 ± 2.4
2 Section of manganite showing maximum A_λ (c) above small negative values of $2\mathcal{V}$ for most wavelengths. Most sections, however, give a positive sign in white light with gypsum plate.
3 F = Faint; D = Distinct; S = Strong; N = Negligible.

RESULTS OF MEASUREMENTS

Maximum corrected values of A_r and corresponding values of 2ϑ for various minerals are given in Table I. Values for reflectivities, as reported in the literature, are given in the third column from the right. Values of A_r and 2ϑ for a given anisotropic mineral vary with the crystallographic orientation of the section under observation. Only the maximum value of A_r is characteristic of a mineral. If the mineral is uniaxial (hexagonal or tetrago-

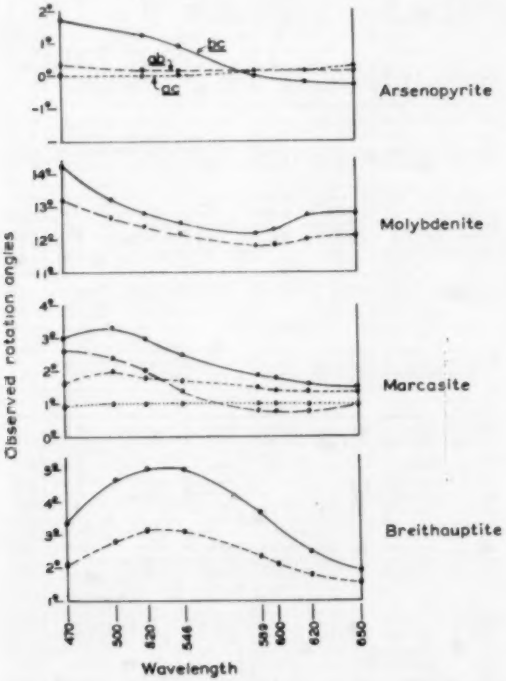

FIG. 2. Dispersion of A_r of four anisotropic minerals.

nal), the maximum and diagnostic value of A_r for all wavelengths of light will be given by a section parallel to the c-axis. If the mineral is one of lower symmetry, then in general there will be some one section that will give the maximum and diagnostic value of A_r for a particular wavelength of light. Depending on the mineral, this section may or may not give the maximum values of A_r for all wavelengths of light. For all minerals of lower symmetry, the grain is found that is so oriented as to give the maximum and diagnostic value of A_r for $\lambda = 589$ mμ (A_{r589}). Values measured in other wavelengths from this same grain are taken as the standard values for the mineral, and

these are the values given in Table I for minerals of lower symmetry. Arsenopyrite (Fig. 2) is a peculiar case. Oriented *ab*, *ac*, and *bc* sections of this mineral have all been studied. The maximum value for 589 mμ (0.15°) can be obtained from either the *ab* section or the *ac* section. Both sections give values for other wavelengths that are all less than 0.4°. The *bc* section gives A_{r589} ±0.0°, but values for other wavelengths range from +1.25° to −0.4°, and grains of this orientation show maximum anisotropism in white light. The values for arsenopyrite given in Table I are for this orientation.

Oriented polished sections have been used for determination of A_r for some minerals, as in the case of arsenopyrite, but in general the maximum values have been found by measuring A_r at $\lambda \pm 589$ mμ for 15 to 20 or more of those grains of each mineral that show strongest anisotropism in polished section. For each mineral, the grain showing the highest value of A_r at 589 mμ has then been used for measurements of A_r and 2ϑ for all wavelengths. It will be obvious from the above account that Table I is based on many thousands of measurements of rotation angles and phase differences.

In Table I, most corrected values of A_r are given to the second place, despite the fact that the error of measurement of A_r is commonly 0.1° to 0.2°. The purpose is to reduce the errors involved in computing observed values of A_r, for a particular microscope, from the corrected values of A_r given in the table. Thus for hollandite, the observed A_r at 546 mμ is 4.3 ± .2° for the microscope used. The correction factor for this microscope at 546 mμ is 0.81, and the true A_r is therefore 3.46°. For a microscope with a correction factor of 0.77 (ratio of corrected angle to observed angle = 5/6.5), the observed value of A_r would be $1.3 \times 3.46° = 4.5° \pm .1°$, whereas if A_r were given in Table I as 3.5°, the value of A_r calculated would be 4.56°, or to the nearest .1°, 4.6°. Beyond their use in computing observed angles, the second place figures have no meaning.

Values of A_r and 2ϑ are functions of the immersion medium. In general, oil immersion increases values of A_r. Maximum values of A_r in oil ($n = 1.515$) at 589 mμ are given for most of the minerals in Table I.

Minerals of doubtful identity have been checked by x-ray diffraction methods.

DISPERSION OF A_r AND 2ϑ

The data of Table I indicate that several types of dispersion of A_r are shown by anisotropic ore minerals (Fig. 2). In the simplest cases, A_r either increases progressively with increase in wavelength (e.g. covellite, klockmannite) or decreases progressively with increase in wavelength (e.g. chalcophanite, niccolite). For some minerals, however, A_r passes through a maximum in the mid-portion of the spectrum (e.g. breithauptite) or a minimum (molybdenite). The former is the more common case.

Since for a given wavelength of light the rotation produced by a mineral is a function of crystallographic orientation, dispersion of A_r is also a function of orientation. If the mineral is uniaxial (hexagonal or tetragonal), the maximum dispersion will be given by a section parallel to the *c*-axis, which

will also give the maximum values of A_r for all wavelengths. Dispersion curves for differently oriented grains of a uniaxial mineral such as molybdenite or breithauptite (Fig. 2) will always have the same form. This may or may not be true for minerals of lower symmetry. The cases of arsenopyrite and marcasite (Fig. 2) are illustrative. For arsenopyrite, A_r passes through zero in a *bc* section at 589 mμ. At a given 45° position of the section, A_r for wavelengths above 589 mμ is, in other words, opposite in sense to A_r for wavelengths below 589 mμ.

Dispersion of 2ϑ likewise exhibits various patterns. For some minerals, such as pyrrhotite, the dispersion is negligible. For others, such as native tellurium and covellite, 2ϑ increases progressively with wavelength. For still others, such as stibnite and freieslebenite, 2ϑ decreases with increasing wavelength and may pass through zero at a particular wavelength (bismuthinite), or may pass through a maximum (cassiterite) or a minimum (tetradymite).

SIGN OF THE PHASE DIFFERENCE IN WHITE LIGHT

In the next to last column of Table I, the sign of the phase difference in white light is given, as determined with the gypsum plate according to the procedure described on a previous page, together with a description of the strength of the color effects on the plate. The notation "nil" in this column indicates that the color effects produced by the mineral are too weak to be useful for determination of the sign.

COLOR PHENOMENA IN POLARIZATION FIGURES

The polarization figures seen when various anisotropic minerals are examined conoscopically in white light include many that are strikingly colored and highly characteristic of particular mineral species (4, Fig. 11, and 5). The vivid blue and red figure of covellite and the figures of arsenopyrite, enargite, rickardite, marcasite, umangite, klockmannite, and niccolite are examples. Descriptions of the colors of the figures for various minerals are not given, however, in Table I. The reason is that the distinctness of colorations, particularly colorations of the figures of minerals of low dispersion, is to some extent a function of the particular optical system and illumination in use. Universally applicable descriptions cannot be given. For any given optical system, however, the effects are constant. The phenomena are best used by the individual observer on a comparative basis, i.e., by comparing the figure of an unknown mineral with those of known minerals, using the same optical system. The figures are then very useful; for example, pyrolusite ($r > v$) can instantly be distinguished from manganite ($v > r$) by inspection of the polarization figure.

The color phenomena are functions of dispersion, and the simpler cases of dispersion can be used for qualitative determination of the dispersion formula (5). Where A_r passes through a maximum or a minimum in the midportion of the spectrum, however, interpretation of color fringes is not always feasible.

EFFECT OF THE SURFACE FILM ON ROTATION PROPERTIES

During the polishing of a mineral grain, a thin film is produced at the surface of the mineral. The film has optical properties different from those of the mineral itself. If the film is thick enough, the properties observed for the grain are altered. This film and its properties have been studied by many investigators. For the present purpose, two aspects of the surface film are important. First, does the surface film have a measurable effect on values of A_r and 2ϑ? Second, regardless of the answer to the first question, do the values of A_r and 2ϑ vary from polished section to polished section?

The first question can be answered only by comparison of values of A_r and 2ϑ given by a cleavage face or crystal face compared with values given by the same cleavage or crystal face after polishing. Several studies of the effect of polishing on A_r (observed angles are referred to in every case below) have been made in our laboratory. J. F. Davis has investigated the effect of polishing on A_r for stibnite, enargite, teallite, and wolframite. Values of A_r for 470, 500, 520, 546, 589, and 620 mμ were measured from a cleavage surface of stibnite. The surface was then polished five successive times on a Graton-Vanderwilt machine. The amounts of abrasive, the dryness of the lap, and the polishing time were varied from run to run. After each polishing run, values of A_r for various wavelengths were redetermined. At 470 mμ the rotation angle was decreased from $3.1° \pm 0.1°$ to $2.6° \pm 0.2°$. Values for other wavelengths showed no changes within the error of measurement ($\pm.2°$). A similar study of enargite (basal pinacoid) showed no measurable change in values of A_r for various wavelengths with successive polishings.

On the other hand, polishing of two (001) cleavage sections of teallite [1] showed that polishing caused a decrease in A_r in all wavelengths, from 2.6° to 2.9° to 2.0° to 2.1°. The change was the same, however, for both specimens.

For wolframite, (010)[1] cleavage faces of three fragments were carefully oriented and mounted in a bakelite block, and the cleavage face of a fourth fragment was mounted in a second block. Values of A_r in all wavelengths were measured on each cleavage surface before and after polishing. All showed the same decrease in A_r (0.6° to 0.7°) after polishing.

L. T. Larson measured A_{r589} on unpolished prism faces of proustite and then on prism faces after polishing. Values for unpolished faces were consistently 0.6° higher than for the polished faces.

Cameron measured A_{r589} and A_{r546} on (001) cleavage faces (these give maximum A_r) of quenselite. Values obtained were $3.0°_{546}$ and $3.1°_{589}$. After polishing with diamond abrasives, values were $3.1°_{546}$ and $3.3°_{589}$. After harsh, dry polishing values were $1.7°_{546}$ and $1.6°_{589}$. The surface was then buffed for one minute on felt with magnetic rouge; values obtained were $3.3°_{546}$ and $3.3°_{589}$. All values are $\pm 0.2°$. It would appear that, within the limits of error of measurement, no change is produced by normal polishing. The surface film produced by severe polishing is removed by brief buffing.

For pyrrhotite, surfaces prepared from various specimens of ore from

[1] Maximum values of A_r and 2ϑ for teallite and wolframite have not yet been measured, hence these minerals are not included in Table I.

Sudbury, Ontario, polished by several different methods at various times, have been examined. All give the same maximum value of A_r. Two separate polished sections of berthierite were studied statistically by J. F. Davis, who found 5.5° and 5.6° respectively, for the maximum value for A_r at 589 mμ. R. K. Sorem found that a natural cleavage face of groutite gave $A_{r589} = 5.8 \pm .2°$. The same cleavage buffed for four successive 5-minute periods gave, after each buffing, 4.8° ± .2°, and the same value was obtained after repolishing with diamond abrasives.

In determining standard values of A_r for a mineral, we have adopted the practice of buffing the sections, after measurement, for successive 3-minute periods, and measuring the value of A_r at 589 mμ after each buffing. For some minerals, the maximum value of A_r has been determined in each of two or more polished specimens, polished at different times and subjected to varying periods of buffing. For most minerals, polishing either produces no change in values of A_r or it produces a definite change in value that is constant from section to section within the limits of accuracy of the methods of measurement. The values of A_r as measured in polished section can therefore be relied upon for mineral identification. This conclusion is consistent with results of studies (2, 7) on the effects of polishing on values of reflectivity. Values of A_r are functions of the uniradial reflectivities (1) of anisotropic minerals.

Several exceptions to the above statements have been found, however, among the minerals listed in Table I. Hausmannite polished by different techniques and for various times shows an irregular change of the value of A_r within a range of 2°. The values given in Table I are the maximum values observed. Covellite is an extraordinary case. Larson studied a polished section of covellite cut ⊥ (001). After initial preparation, the section showed $A_{r589} = 14.6°$, and this value and values measured for other wavelengths at the same time are given in Table I. The section was then buffed for successive 3-minute periods, and A_r was measured with each buffing. A_{r589} rose irregularly to 23.6° after 39 total minutes of buffing and fluctuated thereafter between 22.8° and 23.6°. Fortunately, the values of A_r for covellite are so high and the other optical properties of the mineral so distinctive, that confusion with other minerals would be difficult. Larson found, in a study of certain anisotropic silver minerals, that pearceite repeatedly polished shows irregular variations in A_{r470} over a range of 1.0°, and in A_{r520} over a range of 1.7°. The value of A_{r589} varied only over a range of 0.4°, against an error of measurement of ±0.2°. Similar results were found for polybasite. Proustite, pyrargyrite, andorite, diaphorite, freieslebenite and miargyrite showed no effects of polishing on A_r.

It seems evident that effects of polishing on values of A_r must be investigated for each mineral. Data thus far obtained indicate, however, that for the majority of minerals the effect of polishing, if any, is consistent from section to section, and that the standard values of A_r given in Table I, except as noted therein, can be relied upon for identification. Experience in our laboratory, where rotation properties have been used and checked over a period of years by a number of observers, is in agreement with this conclusion.

The effects of polishing on 2ϑ have not yet been appraised. For this reason, and because accurate setting of the mica plate is often difficult, we use values of 2ϑ only in a qualitative way in identification.

APPRAISAL OF RESULTS

From the data presented in this report, and studies on which the data are based, certain conclusions are drawn:

1. The range of rotation angles of anisotropic ore minerals in air is very broad relative to the accuracy of measurement, which in the usual case is $\pm 0.1°$ to $\pm 0.2°$. Measurement of rotation angles is therefore a useful aid in mineral identification.
2. Identification is further aided because dispersion of rotation angles with wavelength of light is strong for some minerals but is weak for others, and because several types of dispersion patterns are exhibited by ore minerals.
3. Determination of the sign of the phase difference for a mineral is a supplemental aid to identification.
4. The color phenomena of polarization figures are in some cases distinctive and are very useful in identification of certain minerals.
5. For most of the minerals studied, surface films produced by polishing do not affect the reliability of rotation angles for identification of minerals in polished surfaces. Further studies of the effects of polishing on values of 2ϑ are needed.

Students of ores and ore minerals have long since learned that no one technique nor any one set of properties is completely adequate for mineral identification. Physical properties, qualitative optical properties, reflectivities, chemical properties, and other properties have all been tried. All contribute to identification, but none is alone sufficient. The nearest approach to a self-sufficient means of identification is X-ray study, but even X-ray investigation needs assistance from other techniques in isolating minerals to be photographed and in interpreting photographs of mixtures of minerals. Furthermore, it is time-consuming.

Careful study of Table I will indicate that rotation properties are no panacea for all the problems of mineral identification. Anyone who uses them, however, will find that they are in many cases of substantial assistance, and that their use greatly extends the range of minerals that can be rapidly identified under the microscope without recourse to special techniques. Inasmuch as qualitative observations of anisotropism are by themselves extremely useful in ore mineral identification, it is to be expected that quantitative measurements will prove even more valuable. The data of Table I provide, for the minerals reported, the basis for quantitative study of anisotropism and give definite meaning to effects that we could heretofore describe only by such vague phrases as weak, distinct, or strong.

In Table I, the fundamental measurement is the value of A_r in sodium light. This alone will in many cases suffice for a decision between two possi-

ble minerals. The sign of the phase difference, which can be determined for most of the minerals quickly by use of a gypsum full-wave plate, serves as a further basis of subdivision. The magnitude of the phase difference aids in identification of certain minerals. Finally, the pattern and amount of dispersion of A_r with wavelength may be used and the polarization figures examined. In some cases, the value of A_{r589} in oil immersion will be helpful. Particularly if observations of these properties are combined with measurements of reflectivity and hardness, many minerals otherwise dubiously identified can be determined with assurance. The possibilities of identification by joint use of rotation angles and reflectivities may be seen from Table I, in which reflectivities, where known, are also given. The values are taken largely from Uytenbogaardt (11).

FURTHER INVESTIGATIONS

Investigations in the Laboratory of Economic Geology at the University of Wisconsin are continuing with the aim of determining the rotation properties of all anisotropic minerals. There are numerous special problems involved, of which the principal ones are variations in rotation properties within solid solution series and the possible use of rotation properties in determining optical symmetry. Some work on both problems has already been done but is only a beginning of what will necessarily be a protracted series of investigations.

ACKNOWLEDGMENTS

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UNIV. OF WISCONSIN,
MADISON, WIS.,
July 13, 1960

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RELATIONSHIPS OF GENERAL STRUCTURE TO GOLD MINERALIZATION IN THE KILO AREA (CENTRAL AFRICA)

R. WOODTLI

ABSTRACT

The gold mines of Kilo have been successfully worked since the beginning of the century; placers are now exhausted and output comes from primary deposits. Mineralization occurs in quartz veins occupying shear zones and dislocations associated with an old tangential structure. Young radial structures connected with rift tectonics affects the mineralized belts and dislocate the deposits.

RÉSUMÉ

Découvertes et mises en exploitation au début de ce siècle, les mines de Kilo sont toujours en activité; les gisements détritiques étant pratiquement épuisés, la production provient des gisements primaires; ces derniers sont des filons de quartz sis dans des dislocations et dans des zones de laminage dues à une tectonique tangentielle ancienne. De nombreuses failles radiales liées à la formation des fossés africains contribuent à éparpiller les gisements.

INTRODUCTION

THE Central Africa Kilo-Moto Mines are located on the edge of the Western Rift Valley, in the north-eastern part of the Belgian Congo, near the border with Uganda and Sudan.

Gold was discovered by Australian prospectors, first in the South in 1903, in the Chiefdom of Kilo, and then in the North, in 1905–1906, in the gravels of the Moto Creek; hence, the name of the gold mines. According to others, discovery of gold had been reported previously, at the end of the nineteenth century.

It was alluvial gold and several rich placers that enabled the Belgian Government to create an organization that operated these deposits, which in 1926 became the present company. Production has been continuous ever since. The gold output of the Congo is about one percent of the free world supply and most of it comes from the Kilo-Moto district. By the end of 1950 the area had produced 200,000 kg of fine gold; total production must amount now to about 275,000 kg (8,800,000 oz.). Numerous separate small mines have contributed to this total of which the northern mines (Moto) have yielded about 40 percent and the southern mines (Kilo) 60 percent. At the present time mining is confined to the district of Watsa in the mines of Moto, and to the district of Mongbwalu in the south, with some production from Zani in the eastern part of the region. Most of the production comes now from deposits in bed rock and this paper deals with the primary deposits of

the Kilo mines, which extend from 29° 40′ to 30° 30′ E Long. and from 1° 10′ to 2° 30′ N Lat.

Located in the mountainous territory west of Lake Albert, the region of Kilo is rugged with an average elevation ranging between 1,200 and 1,500 m (3,600 to 5,000 feet) above sea level; the peaks of the Blue Mountain Range along the Albert Rift attain a maximum of 8,500 feet; lake Albert lies at 2,000 feet. The large equatorial forest covers a broad area, but most of the country is a savanna, rather arid, with strips of forest along the creeks. Elephant grass grows on the hill slopes, and toward the Sudan border vegetation becomes scarce, consisting mainly of short grass and thorny trees.

The climate is agreeable throughout the year, with diurnal temperatures ranging from 68° to 77° F. The nights are cool; annual rainfall is between 20 inches in the southern savanna (near Irumu) and 65 inches in the forest of the Ituri River, most of which occurs from April to May and from August to November.

GENERAL GEOLOGY

Considered broadly, the region of Kilo is underlain by a huge body of granitoid rocks including large areas of amphibolite and of moderately metamorphosed rocks held to be Precambrian in age and known as the "Système du Kibali" or "Kibalien." These formations are intruded by numerous diabase dikes that fed some basaltic flows found here and there on the surface (1, 2, 5).

In the east the "West Nile System" includes gneisses, micaschists and amphibolite of unknown age; relations between "West Nile" and "Kibalien" are still unknown, but Kibalien seems to be younger. In the south within the Albert trough and within some smaller grabens are sedimentary detritic formations from Permotrias to Recent age.

Because the important ore deposits are found in such rocks and experience indicates that these areas offer the best prospects for further discoveries, detailed studies have been carried out mainly at the border of the granitoid rocks and in the patches of metamorphosed rocks generally considered as having a sedimentary origin (6).

Granitoid rocks include the following types, from the core outward:

1) Alkaline coarse granular granite with microcline, albite, quartz, biotite, with or without muscovite. Hornblende is scarce. Patches of potash granite occur here and there, but this rock is recognizable under the microscope only.

2) Calcoalkaline granite, the chief constituents of which are microcline and oligoclase, with albite, some andesite, quartz and ferromagnesian minerals.

3) Sodagranite with coarse albite, oligoclase and some potash feldspar, quartz, biotite, hornblende and chlorite.

4) A coarse albitite (called "albitite grenue") with little quartz and some ferromagnesian minerals (chiefly hornblende and chlorite); albite is commonly altered to epidote, zoisite, calcite and quartz (saussuritisation); this rock is widely represented.

5) A coarse plagioclasite (a variety of diorite) with or without quartz, oligoclase, andesite and even labradorite, generally with hornblende more or less altered to chlorite.

Along the boundary between granite and surrounding rocks a pyroxene such as augite, pigeonite or hypersthene is sporadically found in coarse albitite and diorite.

This is a schematic view not everywhere present. Also because of weathering and alteration, the geologist is generally not able to identify these rocks in the field and must study them under the microscope; a petrographic laboratory is an indispensable tool in central Africa.

The granitoid rocks are overlain by amphibolite locally as thick as 400 m; this green rock is fine grained with common hornblende and plagioclase; it commonly alters to saussurite. Some patches of microdiorite and gneissic amphibolite are mixed with the normal amphibolite; some bodies of fine-grained albitite occur within the granitoid rocks and the amphibolite. When they lie near the contact between granite and surrounding rocks, they are commonly foliated and gneissic and have the appearance of crystalline schists. The contact seems to be normal, sublevel and gently undulating. In places it looks like a migmatite zone, with alternate bands of granitoid rock and amphibolite or podlike lenses of one rock in the other. "Migmatite" is used here only as a descriptive name and does not refer to the origin of the rocks.

Crowning the series, the "Kibalien" consists of a great variety of schistose and massive epimetamorphic rocks extremely variable in nature. Most abundant are (4):

1) Among the massive rocks—fine and microfine-grained albitite and plagioclasite which, by alteration and weathering, look like schists;

2) Among the foliated rocks—calcareous schist, blackschist, micaschist, chloritoschist, amphiboloschist, talcschist, albitoschist.

A striking feature is the presence of siliceous rocks with a high percentage in magnetite, which are locally known as itabirites; in the Kilo area most of them are rather banded ironstones or jaspilites and taconites. In a few localities in the Kilo region and more widely in the Moto mines are true itabirite, almost monomineralic, consisting preponderantly either of magnetite or of hematite. Owing to rapid variation along the strike and with depth, and through lack of fossils and good markers, it is impossible to establish a useful general succession or zoneography of the Kibalien.

Numerous basaltic dikes are intruded through all of these rocks. They are quite common in composition, like a plateau basalt, with a pyroxene (generally pigeonite) and labradorite-bytownite (subsidiarily quartz, apatite, ilmenite). They are coarse to fine grained, in places glassy and subjected to saussuritisation and uralitisation. Diabase dikes are well represented throughout the Kilo area but are particularly abundant near the Western Rift and decrease in number towards the North-West.

To summarize, the sequence met in the area is a granitoid basement overlain by amphibolite and by Kibalien, all these formations being intruded by diabase dikes. This broad zoneography is shown in Figure 1.

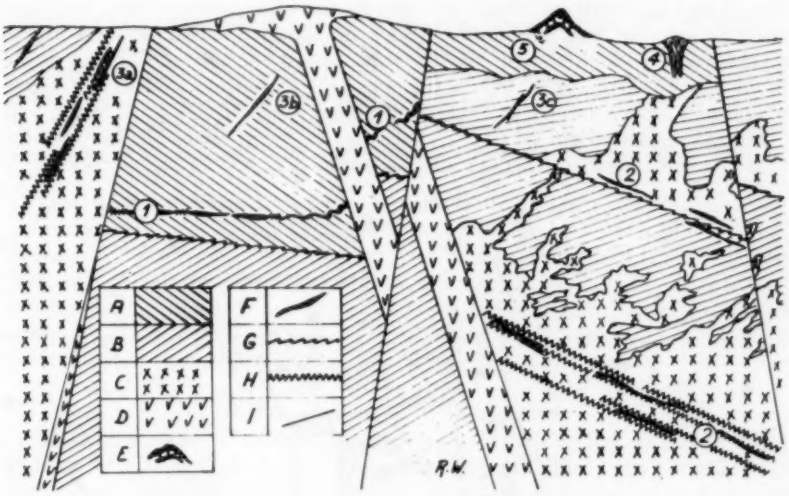

FIG. 1. Diagrammatic geological section. A = Kibalien, B = Amphibolite, C = Granitoid rocks, D = Diabase dike (or dolerite), E = Itabirite, F = Quartz vein, G = Reverse fault, H = Shear zone, I = Radial fault. 1, 2, 3, 4, 5 = types of ore-bodies mentioned in text.

It must be added that commonly all of these rocks, except the diabase and the itabirite, are deeply weathered. The writer remembers drillings, at the bottom of a valley, which failed to supply recognizable rock cuttings before a depth of 300 feet. Consequently geological mapping is often very slow.

STRUCTURE

The general structure will be discussed under two headings to conform with the major periods of structural activity, namely, old tangential tectonics, and young radial tectonics (5).

The tangential structure is difficult to study but is economically important. It is the writer's opinion that the primary gold deposits of the Kilo region are basically controlled by old structures. Consequently, the search for new deposits, particularly for hidden deposits, must be preceded by detailed structural studies in the field. Until now few opportunities to observe the old structure were available except in the ore-bodies or close to them. These tectonics are characterized by faults and shear zones dipping less than 30°. In some cases the Kibalien seems to be thrust over the amphibolite, but this is difficult to prove. The outcrops are generally too deeply weathered or altered to provide decisive evidence and there are very few underground workings within that contact. Only drill cores supply evidence. Ordinarily they indicate an abrupt change from Kibalien to amphibolite. They intersect, for example, a talcschist with a high proportion of carbonate,

then 1 to 3 m of actinoloschist, and then amphibolite carrying basic feldspars such as andesite and labradorite. Such an abrupt change between highly metamorphic rocks and only slightly transformed carbonaceous rocks seems to be hardly explainable by continuous metamorphism. Moreover, actinoloschist is known to appear along fault planes, walls of dikes and veins, and other mechanical contacts (3). Therefore, it seems plausible to conclude that in some instances Kibalien may be thrust over the basement rocks.

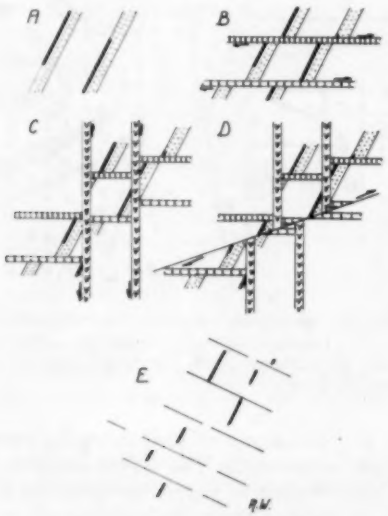

FIG. 2. Influence of radial tectonic on the distribution pattern of ore-bodies. A, Any structure with ore-deposits in black; B, First breaking phase E-W, small offsetting and filling by some vein material; C, Fissuring N–S, important horizontal displacement and intrusion of diabase; D, New breaking WSW–ENE, horizontal displacement and mylonitization along the fault plane without intrusion; E, Distribution pattern of the ore-bodies. Without geological mapping faults WNW–ESE could be assumed.

Good examples of tangential tectonics are found along the border of a granite area in the district of Mongbwalu. Here shear zones and ribbons of mylonite clearly indicate reverse faults dipping about 30° toward the granite. The surface expression of such faults is a shattered zone in laminated or in highly micaceous granite. In hard rocks well marked cataclastic and banded structures are sometimes found; such rocks are sometimes called "migmatite" which is a good descriptive name. The general trend of the Kibalien foliation and local folding also seem to be controlled by the old structure.

Much more is known about the radial tectonics. They are clearly and closely related to rift faulting and to basaltic intrusions, and much information has been gathered about strikes and dips of faults and dikes, relations between various formations, and about old and young grabens. It seems that these

structures were initiated very early, perhaps in Permian time or even earlier, and they are still active.

Through lack of time markers it is difficult to establish a good chronology of the events. Only within the Albert Rift are some beds dated by fossils; they are scarce and insufficiently studied. Rift faulting does not show direct connection with gold deposition but contributes to the shattering of the ore bodies (Fig. 2).

Directions of radial faults and of diabase dikes are extremely variable, but they are apparently always related to known grabens or horsts of great magnitude. Strike-slip displacement and vertical displacement along the dikes is in general rather moderate, but in a few places is large. Horizontal and vertical displacements along radial faults are important; several hundred meters or even several kilometers of offsetting are common. The hade of these faults is almost always less than 30°.

ORE DEPOSITS

Until now gold is the only economic metal of the Kilo region. Huge high-grade iron deposits are known, but are not yet mined. In the Kilo area the economic gold deposits are restricted to quartz veins, and all of them are related to disturbed, faulted, or sheared rocks. Of course, not all mechanically disturbed rocks are gold-bearing. The grade of gold quartz varies greatly from one deposit to another and within the confines of any particular body. Ore shoots are controlled by secondary local structures, generally younger than the main features, and physical characteristics of the rocks seem to be more important than the chemical ones. However, it must be pointed out that in all cases the surrounding rocks have a high content of albite, but soda feldspar is so widespread that this does not serve as a guide for prospecting (3).

The following development refers to the diagrammatic section of Figure 1. The central part of the diagram represents the large mine of Senzere near Mongbwalu. The Kibalien, called here "upper block," is a foliated mass of highly carbonaceous talcschist with layers of fine-grained albitite. It rests on the amphibolite or "lower block" and the contact between them is considered to be a thrust plane. A large vein of quartz (No. 1 of the diagram) is roughly parallel to that contact 20 to 30 m above it and seems to be located in a shattered zone. The veins are made up of a massive white to gray quartz that in places has been fractured and mineralized with some sulfide (mostly pyrite and pyrrhotite) and gold, which is sometimes visible. Metallic minerals make up only a very small part of the veins and there is no close relation between sulfide content and grade of gold. The complete independence between sulfide and gold is a characteristic feature of most of the Kilo deposits. The case under discussion is a replacement vein and commonly the quartz encloses strips or enclosures of country rocks. In places the quartz fades out, but the trace of a plane of disturbance remains well marked and clearly recognizable. Some meters farther quartz appears again. Near the vein the surrounding rocks are folded and laminated. These facts are well shown by the albitised rocks.

Within the amphibolite and the underlying granitoid rocks, drillings have intersected other dislocations (shear zones and mylonites) parallel to the contact between "upper" and "lower blocks." These findings support the hypothesis that this quartz vein, which extends over a large area, is localized in a secondary flat fault due to thrusting.

Several diabase dikes and subvertical faults have produced many tilted blocks so that the country rock becomes a mosaic. It resembles the breaking up of an ice pack in spring time (5). All of these dikes and faults are post-ore and their result is a shattering and dislocation of the ore-deposits, in plan and in elevation (Fig. 2).

In the right part of Figure 1, is shown a complicated structure due to interferences between old and young tectonics. Rift and block tectonics have superimposed their effects in a region previously cut into subhorizontal slices by tangential tectonics. The upper dislocation is obviously a reverse fault, accompanied by laminated and brecciated rocks; the lower dislocation affects a block without marker beds and expresses itself with mylonite and a shattered zone, so that the whole section is considered a piling of overthrust sheets. Ore-bodies (No. 2) are connected with the shear zones as quartz lenses, quartz veins, and stockworks. The laminated and silicified granitoid country rocks may carry abundant gold, though probably not workable by themselves.

A thorough structural and geological mapping of the district showed the shear zones to have a great lateral extension and to be the host of numerous quartz veins. Some localities were selected for deep drilling and, in the trough of a graben, a thick and rich vein without known outcrops was intersected at an average depth of 450 feet below the surface. This is an example of an unusual case of a blind or hidden gold deposit discovered by geological methods and proved by precisely located bore-holes. It is reasonable to expect other ore-shoots in these shear-zones. It is not clear whether there are one or several major shear zones that have been cut by faulting into several segments that overlap or whether there are various different but related shear zones.

The case of the vein 3a, in the left side of Figure 1, is a common type of deposit, nearly vertical quartz veins in a somewhat disturbed or laminated country rock. In this example the veins lie in a surrounding laminated coarse albitite near the contact with amphibolite; in fact, such gold-bearing veins are known in Kibalien schists (3b) and also in amphibolite (3c) far from any contact with granitoid rocks. One cannot conclude that the emplacement of granite caused the gold mineralization. The feature common to most cases seems to be the presence of an old shearing, which generally affected granite long after its "intrusion."

Ore-body No. 4 (right part of the section) is a rather unusual occurrence, interest of which lies in that it is a bridge between gold-bearing quartz veins and gold impregnations. Here the deposit is a large vein of generally fine, in places microfine, albitite with cores of coarse albitite (or soda-granite). Ore material consists of the albitite with from 20 to 50 percent quartz as stringers and lenses. Contrary to the previous examples of gold deposits,

in the case under discussion gold is found evenly in quartz and in albitite so that the whole mass may be mined with profit. By weathering, this rock gives kaolin mixed with quartz and a high content of free gold.

The four types of gold ore-deposits just reviewed are well represented in the Kilo area; and there also occur indications of a fifth type (No. 5), well known in the Moto area where it has the greatest economic importance (7). It seems to have some kind of relationship with the itabirite. The Kibalien schistose rock ($\frac{1}{3}$ carbonate, $\frac{1}{3}$ quartz, $\frac{1}{3}$ albite) surrounding the itabirite may contain gold disseminated apparently at random. Here and there the content of gold is high enough and there is sufficient tonnage of mineralized rock for mining to be profitable. Until now nothing in the structure, mineralogy, or texture, indicates an ore-body; there is no silicification, no change in the habit or abundance of albite and other minerals, and no variation in the grade and composition of sulfides. Only by assaying can gold be detected (3). The writer suspects some relationship between such ore deposits and the regional structure.

For years belts of mineralization pertaining to two main orientations have been suspected by geologists in the Kilo area (1). The trend of the first set would be E-W or ENE-WSW and four alignments are known: 1) Tsi-Nizi-Labo; 2) Sindani-Camp III–Vieux-Kilo–Alosi; 3) Andissa Valley; 4) Lodjo Valley.

In the writer's opinion they are connected with old Kibalien features and are underlain by siliceous banded ironstones, blackschists, and laminated rocks and they fall chiefly in the 3rd class of the diagram.

The other set is roughly oriented north to south with a low angle of dip to the east and is well known in the region of Mongbwalu. Ore deposits pertaining to class No. 2 of the diagram are related to shear zones in granitoid rocks, and they have such an orientation.

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STATE COLLEGE, PENNSYLVANIA,
July 27, 1960

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MINERALOGY OF THE MA ON SHAN IRON MINE, HONG KONG

S. G. DAVIS

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ABSTRACT

The Ma On Shan Iron Mine in the Leased Territories of Hong Kong has a proved reserve of 10 million tons of magnetite. The mine, which until 1957 was worked open-cast, is about 250 m above sea level in an area of rugged topography. There are about 2,000 m of underground workings. The production of ore in 1959 was 119,800 tons.

The major rock groups exposed are granites and volcanics that range in age from middle Jurassic to late Cretaceous. The magnetite ore body is lens-shaped and dips gently to the northwest. These rocks have been affected by major earth movements and volcanism during early Tertiary times.

The typical mineral association of the main ore body is magnetite and chondrodite. Surrounding the magnetite is a skarn with a mineral assemblage that includes talc, serpentine, garnet, olivine, pyrite, chalcopyrite and fluorite. At the base of the skarn there are beds of calcite, dolomite and quartzite. The ore body is highly disseminated. A ball-mill using a wet magnetic separator beneficiates it to 56% Fe. Almost the entire output goes to Japan. Quartz veins in the country rocks carry wolframite and molybdenum. Some oxidation and supergene enrichment are present. The mineral assemblage of the skarn in this deposit suggests a contact pyrometasomatic origin. The mineral chondrodite is typical of such a deposit and its occurrence explains why the content of magnesia is so high in the iron ore.

INTRODUCTION

THE following study expands the description of the iron deposit at Ma On Shan recorded in "The Geology of Hong Kong" by S. G. Davis (1).

The views expressed in the earlier description of the mode of occurrence have been considerably modified. This has been made possible by the greater number of new sections exposed by more intensive open cut and underground mining and the very thorough exploration drilling of the Nittetsu Mining Company of Japan.

Location and Topography.—The Ma On Shan Iron Mine is situated in the Leased Territories of Hong Kong about 250 m above sea level. It is on a deeply dissected north pointing peninsula that is surrounded on the west, north and east by a highly indented coastline. It is about 1,300 m southwest of the hill (700 m) also named Ma On Shan. It is also approxi-

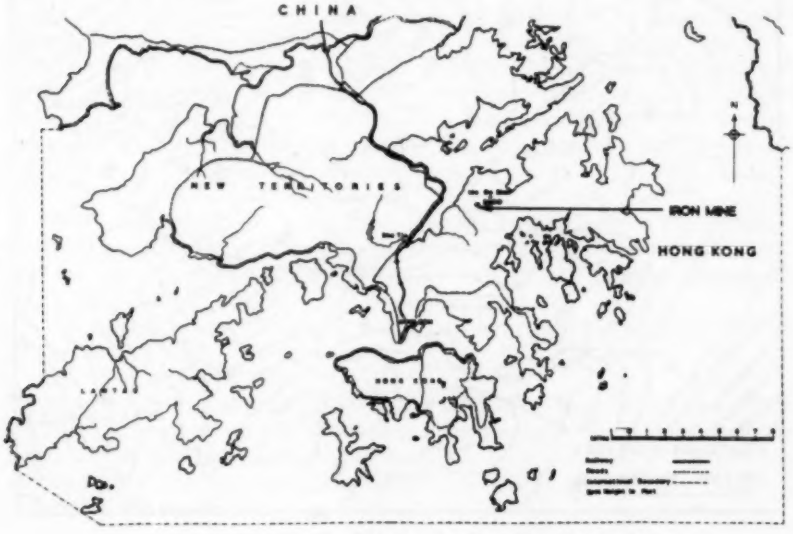


FIG. 1. Location of Ma On Shan iron mine.

mately 16 kms due north of the east end of the Harbour of Hong Kong (Fig. 1). The extent of the horse-shoe shaped open-cut terraces is shown in Figure 7. The length along the north bench is about 500 m and the width 200 m. The long axis of the concealed magnetite ore body is 450 m.

The topography of the mine district is steep and rugged (Fig. 3). Uplands and sharp peaks are the eroded remnants of one small part of the great southeast China batholith of acid rocks. The major rock formations are clearly distinguished by their differing responses to a weathering predominantly controlled by the chemical action of a humid tropical climate aided by typhonic winds.

The mine is linked with Tide Cove, a sea inlet, by a winding but well graded, double-tracked road 2½ kms long. The ore is carried by trucks down the road to a dressing plant which is about 200 m from the sea shore. After

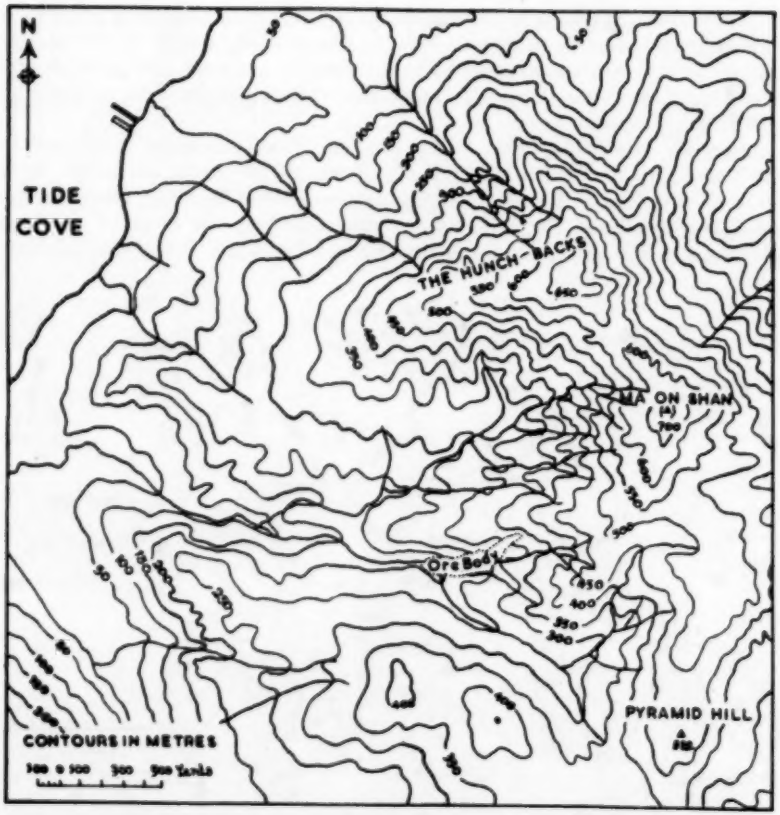


FIG. 2. Location map of the Ma On Shan iron mine.

FIG. 3. Panorama of iron mine looking north. Ma On Shan (Saddle Mountain) in center.

beneficiation the ore is loaded on to lighters and carried to ocean-going freighters that lie about one kilometer off-shore.

History.—The first license to prospect and later mine the Ma On Shan deposit was issued to the Hong Kong Iron Mining Company in 1906. In 1931, the New Territories Iron Mining Company, was granted a Crown Lease for fifty years. In 1940 another company, South China Iron Smelters Ltd., took over the property with the intention of building a blast furnace on the sea front. From 1942 to 1944 the mine was operated sporadically by the Japanese and employed about 1,500 workers. As there was no mechanical transport the ore had to be carried by coolies down to the sea-shore. In 1949 the Mutual Trust Company took over the mining rights. The Company signed a contract with the Supreme Command of Allied Headquarters in Japan to supply 150,000 tons of ore annually. In 1953 the Nittetsu Mining Company of Japan joined with the Mutual Trust Company to operate the mine. During this year the change-over from opencast to underground mining was begun. In March 1954 a fully mechanized ore-dressing plant was installed. In 1959 all mining was underground. Recent annual productions of iron ore for 1957, 1958 and 1959 were 94,181, 105,125 and 119,892 tons.

Mining operations follow the old established Chinese custom of subletting work. The work in 1959 was sub-contracted out to 9 companies who between them employed about 350 miners underground. In addition there was a staff of about 800 employed on the surface.

The average grade of the crude magnetite ore in 1958 had an Fe content of 32 percent. This was beneficiated up to 56% Fe. The dressing plant has a daily capacity of 700 tons.

Reserves are estimated at 10,000,000 tons and planned annual production for the next few years is 120,000 tons of concentrates. Since 1954 the entire output of the mine has been shipped to Japan. Technical assistance and advice in the mine and dressing plant is provided by a Japanese firm. Diamond drilling is still continuing in order to establish the reserves accurately.

PREVIOUS GEOLOGICAL WORK

Only one brief report on the deposit by C. M. Weld (6) in 1914 has been published. Several private reports and maps were in existence before World War II but were lost during the Japanese occupation. Mine maps prepared since 1953 by Japanese engineers together with some general information from drill logs have been made available by the Mutual Trust Company.

ROCK FORMATIONS

Extreme tropical weathering, coupled with the major erosion of the Tai Hang river and its tributaries which have cut steep-sided valleys, has been responsible for exposing the rock sequence. Thus the contact between the Repulse Bay volcanics (Middle Jurassic) and the Hong Kong granite (Upper Cretaceous) is clearly exposed (Fig. 4). One kilometer to the west of this contact is a long narrow outcrop of the Tolo Channel sediments (Lower Lias)

which are marine in origin and are regarded as the basement beds of the area. Three kilometers to the north are Pat Sin conglomerates (Lias) which are of continental origin and derived from the earlier basement formation of the Lower Lias.

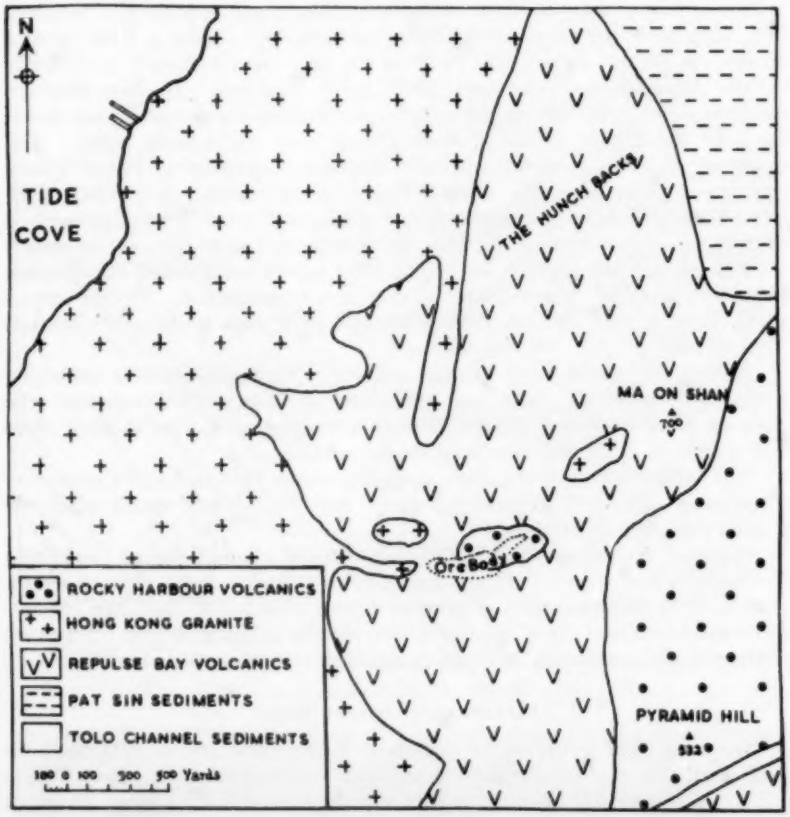


FIG. 4. Map of Ma On Shan District, New Territories, Hong Kong.

About 18 kms due north of the iron mine is an agglomerate that is included in the Rocky Harbour volcanics (Miocene) and contains boulders of limestone. This agglomerate rests on a felsite surface which dips north at an angle of 45°. The agglomerate appears to be the filling of a volcanic fissure. The limestone boulders on the evidence of their fossils have been classified as Permian in age. It is suggested that this part of the New Territories of Hong Kong is underlain by late Palaezoic formation from which fragments in the form of xenoliths have been brought to the surface during

the intense diastrophism and volcanic eruptions of the Jurasside Revolution (Yen Shan Number One).

In the immediate area of the iron mine are beds of calcite, dolomite and quartzite. These are certainly not younger than any other rocks in the area and are possibly contemporaneous in age with the limestone in the agglomerate.

Petrology.—The Repulse Bay volcanics are the principal country rocks into which the granite was intruded. They vary widely in character, even within a few yards. The complete range of beds is estimated to have a thickness of over 2,000 feet. The most typical rock-type is a rhyolite porphyry. The groundmass is normally a deep brown color and has a crypto to microcrystalline texture. Flow structure is common. The phenocrysts are alkalic and sodic feldspars and quartz of about 1 to 1.5 mm.

The rocks that form the immediate roof of the granite intrusion show strong contact metamorphism. Commonly they are altered to a quartz sericite rock, and in places are highly silicic and cherty. Andalusite crystals, up to three centimetres in length, are common.

The typical Hong Kong granite is a coarse, equigranular rock with an average grain size of about 2 mm. Some pink feldspars occur up to 1 cm in length. The rock is light in color with a white or pink tint according to the color of the feldspar. The essential minerals, quartz and orthoclase and plagioclase feldspars are easily distinguished by eye, as is the biotite, which is commonly present. Other minerals present in different specimens are magnetite, zircon, hornblende and muscovite.

STRUCTURE

The violence caused by the intrusion of the granite into the volcanics is recorded by major faulting and fracturing. Along the north face of the 1st Bench there are two parallel faults well marked by slickensides. They can also be distinguished, together with a third fault, just above the 5th Bench. They strike approximately N 70° E. The north tributary of the west-flowing Tai Hang river exploited the weakness of a fault zone above the iron ore mass. The down-cutting of its valley later exposed the deposit.

During heavy typhonic rains the loose cap or mantle-rock above the 300 m level has slumped several times to cause serious landslides. The mantle-rock is a mixture of lateritic clay and rock boulders and fragments. Landslides occurred in 1957, 1958 and June 1959. On the last occasion the slump was between two major fault zones and completely blocked the portal to the 240 m level.

Post-mineral movement is clearly seen in the 1st and 2nd sub-level crosscuts of the main ore body. One major fault, that is primarily lateral, trends N 25° W. Other structural evidence elsewhere in Hong Kong and Kwangtung suggests that this post mineral faulting is related to the Pliocene diastrophism of the Alpine revolution.

ORE DEPOSIT

In Hong Kong, as in the neighboring province of Kwangtung, the metamorphic contacts of the granite with the volcanics are not generally accom-

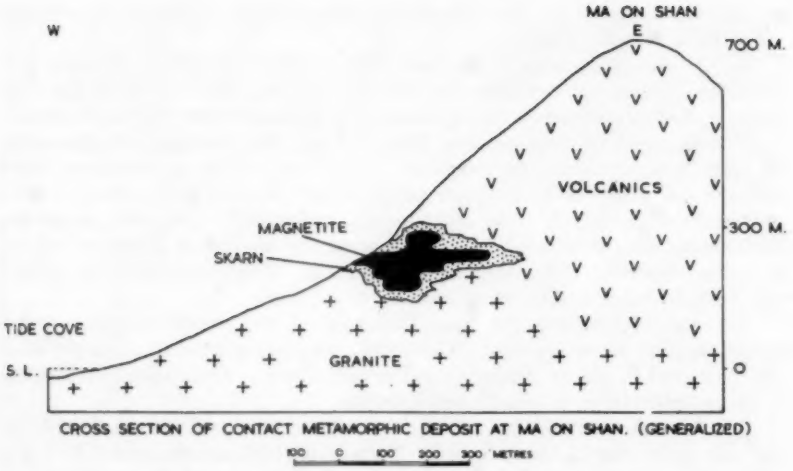


FIG. 5.

panied by mineral concentration and deposition, but involved only recrystallization of the country rock to give hornfels with a granular fabric. However the contact metamorphism at Ma On Shan is distinguished by the magnetite deposit (Figs. 5, 6). The genetic relationship of the ore body to the granitic intrusion (Figs. 5, 8) points clearly to one of magmatic emanations given off from the magma during its consolidation. A magnesium limestone provided favorable fluxing conditions for pneumatolytic transfer and the concentration of the iron oxide.

The deposit is broadly divisible into two members: a central lenticular mass of magnetite which is intermixed with hematite; and a surrounding calc-silicate skarn or tactite.

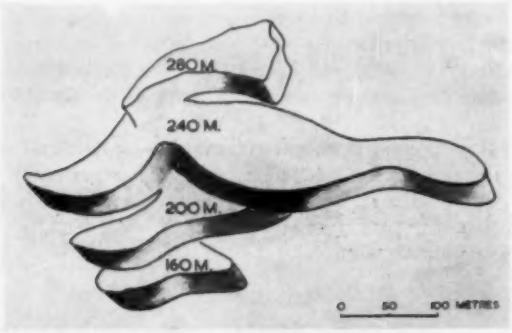
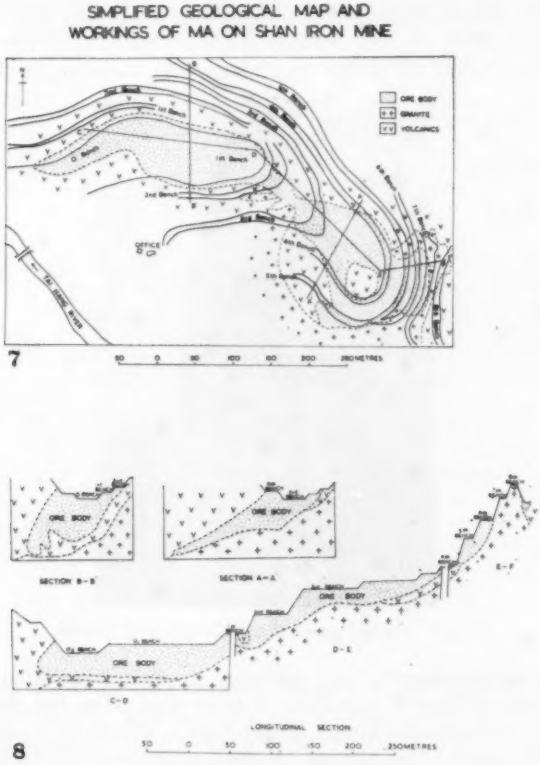


FIG. 6.

Magnetite.—The magnetite from a mining point of view is capricious since it is highly disseminated. It occurs in bodies ranging from minute isolated crystals that are cemented or held together by silicates up to masses of many tons. In its pure massive form the magnetite is black, magnetic, rather soft and granular. There are no well defined boundaries between the ore and



FIGS. 7, 8. Plan and sections of ore body.

the gangue. The ore grades imperceptibly into the matrix, which is within the range of a granite-granodiorite-quartz group. Some parts of the matrix are pegmatitic with well developed clusters of quartz, feldspar and biotite.

Hand picked lump ore runs about 47% Fe, with about 13% SiO_2 and MgO each; concentrates run around 57% Fe, 8% SiO_2 and 6% MgO.

Skarn.—A calc-silicate-hornfels rock is characteristic of both the hangingwall and footwall of the main iron mass. The contact aureole or skarn carries diminishing amounts of magnetite, hematite and other iron minerals, especially chondrodite. The hangingwall section exposed along the 5th Bench

of the open cut has a garnet-pyroxene-epidote zone of high-temperature minerals. Some clusters of large garnets (Fig. 9) make up between half to three-quarters of the total volume of the rock. Also present in the footwall along the Zero Bench of the open cut are small patches of coarse crystalline pinkish wollastonite. Pyrite, chalcopyrite, and galena are also present in minor quantities. Cores drilled from the footwall at depths of about 40 m have revealed some beds of calcite, one twenty feet thick. Small vugs yield beautiful crystal specimens of both nail head and dog tooth calcite and blue to purple clusters of cubic fluorite.

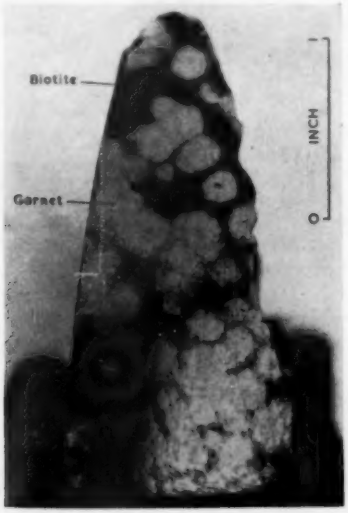


FIG. 9. Hand specimen. Garnet and biotite.

No clear cut metamorphic rock zones are traceable away from the contact. However deeper drilling to distances of more than 80 m has revealed some grading into a chondrodite-quartz-calcite zone. Traces of brownish rhodochrosite are thought to be present in small quantities. Isolated patches of yellowish and brownish green andradite garnet are associated with sericitized feldspar and partly altered chondrodite. Relics of the original volcanic country rock occur commonly in the new hornfels structure. Talc and serpentine are present in masses and also as veining through quartz layers. A one-meter bed of quartzite shows under the microscope a mosaic of interlocking quartz crystals together with small clusters of magnetite and small amounts of green biotite. Also associated with the calcite are wollastonite, olivine, and green-white ophicalcite.

At a distance of about 100 m from the main contact in the hangingwall, there are radiating clusters of slender tourmaline embedded in quartz. At this distance pockets of white kaolin and greisen are common. The mica of

the greisen carries some partially altered feldspar. Quartz veins and stringers containing wolframite and molybdenite occur irregularly. The wolfram is present in sufficient economic quantities for it to be worked by the iron miners in their free time.

Paragenesis.—The order of crystallization in the skarn, based on thin sections and polished sections, is shown graphically below. The deposition of the quartz, pyrite, chlorite and fluorite are presumed to have spread over a longer period than both the earlier high-temperature minerals and the later low-temperature minerals that crystallized below the critical temperature of water.

Origin of the Ore.—The various features of the mineral assemblage of the skarn have already been indicated. The origin of the ore is contact metasomatic. Particularly the mineral chondrodite is typical of deposits of this kind. Its occurrence explains also why the content of magnesia is so high in the chemical analyses of the ore.

It seems reasonable to assume that there was a pre-existing bed of magnesium limestone at the contact of the granite and volcanics that acted as a flux to concentrate the iron. The age of the deposit is Paleocene to Eocene.

ACKNOWLEDGMENTS

The assistance of Mr. W. K. Yu, Managing Director of the Mutual Trust Company and Mr. M. Nishiwaki of the Nittetsu Mining Company in allowing me access to previous private and confidential information was of great value. Dr. Katsu Kaneko, Director of the Japanese Geological Survey and Mr.

J. H. Knapp, Government Superintendent of Mines provided valuable chemical and statistical details. Mr. Harold Service, late Director of Geological Survey, Malaya, following a visit to the mine gave suggestions on the mineral assemblage as shown to him. Mr. David C. Y. Lai (5) of the Department of Geography and Geology, University of Hong Kong took many of the pictures and gathered background material.

UNIVERSITY OF HONG KONG,
June 15, 1960

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GOLD IMPREGNATION DEPOSITS IN THE MOTO AREA (CENTRAL AFRICA)

R. WOODTLI

ABSTRACT

A short description is given of impregnation deposits of the Moto gold field in the NE part of the Republic of Congo. These ore-bodies are located in a Precambrian schistose rock containing carbonate, quartz, and albite due to metasomatic processes. There is no correlation between quartz stringers, abundance and composition of sulfide and gold grade. A low grade halo furnishes a valuable guide in the exploration for new ore bodies. Other significant features include a remarkable parallelism between the ore-bodies (direction NNE, dip 24°) conformity to rock foliation, occurrence of itabirite and black schists, and some lenticular sulfide bodies.

RÉSUMÉ

L'auteur décrit sommairement les gisements aurifères d'imprégnation de la région de Moto au NE du Congo. Ces gîtes sont enchâssés dans des roches précambriennes métamorphiques à carbonate, albite et quartz d'origine métasomatique. On ne trouve pas de corrélation entre le quartz filonien, l'abondance ou la composition des sulfures et la richesse du minerai. Des auréoles de roche à basse teneur servent d'indicateur dans la recherche de nouveaux gisements. Un trait structural important est le parallélisme remarquable présenté par les gisements dont l'axe est quasi invariablement orienté au NNE, avec un plongement moyen de 24°; on note aussi qu'ils sont en concordance avec la schistosité. Des bancs d'itabirite et des lentilles de schistes noirs apparaissent souvent dans le voisinage des gisements.

In another paper (4) a general view of the gold deposits in the Kilo area was presented, largely based on the writer's own field work and detailed mapping. These ore-bodies are worked by the Kilo-Moto Gold Mines Corporation, a large company operating in the north-eastern part of the Republic of Congo. This paper is concerned with some aspects of the mineralization in the northern gold field, known as the Moto area.

GEOLOGIC SETTING

The geological setting in the Moto district is generally the same as in the Kilo area; in both regions a granitoid basement is overlain by amphibolite and by moderately metamorphosed rocks considered to be precambrian in age and known as the "Système du Kibali" or "Kibalien". Whereas diabase dikes are very abundant in the Kilo area they are scarce in the Moto region except in the southern part. The problem of the relations between "Kibalien" and "Western Nile System" remain unsolved. The Western Nile, previously

called "Archean," consists mainly of gneisses, gneissic amphibolite, mica-schist and quartzite. In places the Kibalien seems to overlay the Western Nile formation with an angular unconformity, and in other places the rocks of the two systems seem to merge into one another. Nothing is known about the relations of the granite masses connected with each series of rocks. Also much of the country has been deeply eroded and reduced to a partial peneplain with huge remnant mountains. Because of this peneplanation, the weathered layer is thick. There are very few outcrops and geological mapping is hindered in the lower part of the country. Many hills rising from the general level of the Kibali River peneplain are unsymmetrical ridges with a steep southern slope and a long northern one generally parallel to some resistant itabirite layers. They may be crowned by laterite crusts. Although very little is known about the structure, it is apparent that some morphological features are closely related to the radial tectonics. But that dependence is less conspicuous than in the Kilo district, which lies on the edge of the Western Rift Valley.

ROCK FORMATIONS

The gold quartz vein deposits of the various types known in the Kilo (4) area occur and are mined in the Moto gold field also, but they have no great economic importance. Most of the production comes from gold impregnation ore bodies in the Kibalien schistose rocks. The following is the first description of these deposits in English.

North of the Watsa station a long section of the Kibali River is cut into a particular geologic formation confined to an area of about 1000 km^2; it is considered to be a facies of the "Séries du Kibali." The main components of that rock are carbonate, albite, and quartz, each making up more than 30 percent. Subordinate minerals are sericite, chlorite, epidote, tourmaline, iron oxides, sulfides and a little graphite. Microscope study shows that the proportion and facies of the minerals are extremely variable.

Carbonate.—The most common mineral is a ferroan variety of dolomite—*ankerite*. In some exposures *siderite* occurs and may form large cores of siderite marble, generally more or less colored by various minerals, the principal ones being magnetite, hematite, and iron sulfide. Ankerite and siderite commonly occur together and may be distinguished by their indexes of refraction (under the microscope) or by their alteration, the siderite becoming red brown by oxidation. Calcite is widespread but occurs only in minute cavities and veinlets. It is generally not observable megascopically and its presence is not always ascertained by acid response; the carbonate content of the rock may vary from 0 to 95 percent.

Quartz.—Generally most of the quartz grains are fine and equidimensional, some are disseminated and some form lenses or thin layers of quartzite. Proportion of free silica varies from 0 to 100 percent.

Albite.—The soda feldspar is widespread throughout the Kibalien rocks and occurs either in large district crystals with albite twins (porphyroblasts), either in lath-shaped and twinned sections, or in small nontwinned grains less than fifty microns in diameter.

These three main minerals may be associated in any proportion so that, at least on a microscopic scale, a great variety of rocks occur: dolomite and siderite marbles, silicified marble, quartzite, coarse or fine grained albitites, silicified albitite. In places the rock may look like a granite, or like a lava. Generally sericite and chlorite are common companions of the chief components and the rock may grade into micaschist and chloritoschist so that no definite line can be drawn in the field between these various facies of metamorphic rocks. The specific rock names mentioned above are used only to indicate what is observed under the microscope. Megascopically the rock has essentially a gneissic texture and consists of layers of mingled quartz, carbonate, and albite in any proportion, which are separated by layers of sericite and chlorite. Where the amount of phyllosilicates is small, the gneissic aspect is less evident. Generally these layers are rather thin, short, and lenticular. They may be parallel for a long distance, but generally they are curved, folded, and contorted with numerous small faults. The lenticular texture may be so pronounced that it might be called augen texture. In some cases the rock looks like a breccia with angular coarse or fine fragments. Its color varies from light gray, white, to dark green, with brown shades depending on the abundance of limonite due to oxidation of ferruginous minerals.

No general geologic succession or marker bed has been found within the Kibalien schists, the thickness of which may be about 2,000 m. However, two typical formations not restricted to a particular level are readily recognizable—the itabirite and the black schists.

The itabirite shows a wide range in composition from silicified banded ironstones, such as jaspilites and magnetite quartzite, to hematite-schist or almost pure magnetite masses (1, 3). It is commonly associated with bodies of siderite marble, which is progressively invaded by quartz grains and iron oxide crystals so that the marble grades into itabirites. There are commonly large enclosures of more or less silicified marble in the itabirite and bands or lenticular bodies of iron ore within the carbonaceous rock. Generally itabirite occurs as long ribbons a hundred meters or more in thickness and more than thirty kilometers in length (2).

The black schists are generally micaceous and silicified and colored by graphite (or by tourmaline or pyrite). By oxidation and leaching they lose their graphite and turn white (1). In numerous cases they occur in the vicinity of the itabirite or even mixed with it as lenticular folded masses, which may be from a few meters up to about a hundred meters in thickness; they may spread over several kilometers in extent.

A thorough microscopic investigation of all these rocks has been conducted by Dr. Sorotchinsky (1). According to him, the primary rock was a dolomite (ankerite) marble which has been heavily albitized. Then closely spaced planes of cleavage have developed and the rock has undergone strong silicification and sericitization. Silica and soda would be of surficial and metasomatic origin. Itabirite would form by silicification and oxidation of the siderite marble. Black schists would be due to sericitization of the common rock near the topographic surface below swamps or bogs which would supply the black coaly pigment (2, 3).

GOLD MINERALIZATION

Gold mineralization is localized in these rocks without any apparent control by known structural features, mineralogical composition, or texture of the country rock. The ore is the rock itself impregnated by gold, which is generally not visible and may be detected only by assay. However, by crushing and grinding a rock sample (about 10 kg) in a mortar and then by panning the powder, it is often possible to gain a good estimate of the gold content of the rock. This procedure is widely used by prospectors, providing them with a quick field method for determining the value of a prospect. Abundance and composition of the sulfides (chiefly pyrite and pyrrhotite) and abundance of iron oxide (magnetite) has no apparent connection with the grade of the ore (1). The mineralization always is concordant to schistosity or foliation of the country rock and that seems to point to permeability (due to the high proportion of carbonate rather than to sets of fractures) as one of the controlling factors of the gold impregnation. However, the axis of the ore-bodies may not in all cases correspond to the direction of the dip of the foliation. It is observed that the ore-bodies plunge about 24° NNE, whereas the strike of the foliation varies from an E–W to a N–S direction. The foliation dips northward or eastward from 20° to 30°. The parallelism of the ore-deposits is an outstanding feature that has not yet been satisfactorily explained.

Of particular interest are the deposits of *Agbarabo* and *Gorumbwa*. The former has the shape of a gently inclined chimney (24° NNE) the vertical cross section of which is oval. The vertical axis of the oval is about 20 to 25 m long and the horizontal one about 15 to 20 m. In the geologic literature, such shaped ore-bodies are sometimes called mantos. The country rock is the light gray carbonate-quartz-albite schist described above. Within the ore-body the gold content may be as high as 600 g/T. Outward from the rich core, are irregular elliptical halos of low-grade ore, decreasing in value. These aureoles surrounding the main ore-bodies and showing low-grade mineralization over a wide area provide a useful guide to prospectors who will search for gold by first seeking the enveloping halo. On its eastern side the Agbarabo chimney is roughly limited by a large body of siderite marble, quartzite, and itabirite inextricably intermingled. The irregular contact between the schistose rock and this body is probably not a fault plane but it might have controlled the mineralization, possibly by acting as an impervious barrier, since it has been noticed that the low-grade halo hardly extends within the siderite body. The Agbarabo deposit shows some similarities in shape, composition and geological setting with that of *Morro Velho* in Brazil but it must be stressed that in the African deposit sulfide and quartz stringers do not carry much gold. They may be somewhat auriferous but the rock itself yields the wealth, not the sulfide, nor the quartz stringers.

In the Gorumbwa mine the mineralized body, determined only by assay, has exactly the same aspect as the country rock. It is roughly tabular in shape, approximately 70 to 130 m long and 3 to 30 m thick. The gold content may be as high as 300 gr/T. It dips 24° NNE and is concordant to the

foliation. It extends to at least the 500 m level. At the present time, as in other cases, no correlation has been found between either rock composition, texture, or structure and mineralization.

A number of smaller ore bodies and surficial indications of primary gold are known in the Moto district. A common feature of many of them is their association with the carbonate-albite-quartz schist. In general the strike of their axis is NNE and they plunge 24° NE regardless of the foliation strike and dip, but the ore-bodies remain within the plane of foliation and do not intersect it. Generally they show a close association with itabirite bodies or black-schists bands. Both these rocks may carry gold and are sometimes mined with profit, but they are not the favourite host rocks. In a few cases some sulfide bodies occur and are known to be aurifereous. At present they do not appear to be of great commercial value within the Moto gold field.

STATE COLLEGE, PA.,
Aug. 8, 1960

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The author wishes to thank Messrs. L. Bruneel and A. Solvyns of the Board of the Kilo-Moto Gold Mining Corporation who kindly gave permission to publish this paper.

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SCIENTIFIC COMMUNICATIONS

THE CONTENT OF TRACE AMOUNTS OF SILVER IN GALENA ORES FROM BROKEN HILL, AUSTRALIA

E. J. BOUNSALL

ABSTRACT

A quantitative method of chemical analysis for the determination of silver in lead ores has been developed and applied to samples from the Broken Hill region of Australia, whose lead isotope compositions had been previously determined. The results of the analyses do not indicate any significant difference in silver content that can be correlated with differences in isotopic composition or the type of lead ore.

INTRODUCTION

The spectrographic method of analysis for trace metals has certain limitations; its almost complete insensitivity to some elements, and its oversensitivity for others. Where these limitations are not present or where it is essential to analyze large numbers of samples, the spectrographic method is undoubtedly one of the best available. However, if small differences in trace metal content must be detected, particularly at low levels of concentration, quantitative chemical analyses must usually be employed.

A particular example is the determination of trace amounts of silver in lead sulfide (galena) ores. Wasserstein (1) in describing spectrographic analyses of galenas, reported that the role of silver could not be accurately assessed as its sensitivity in the spectrographic examination did not allow the estimation of even approximate amounts.

The necessity of making accurate measurements of trace silver contents of galenas has arisen recently in connection with studies of the variation of the isotopic abundances of lead in lead minerals. Cahen et al. (2) have shown that those ores whose "model" ages (determined from their lead isotope compositions) are older than the ages of the rocks in which they lie ("B-type" leads), have lower trace silver contents than lead minerals whose "model" ages conform to those of their host rocks ("normal" leads). The "B-type" lead minerals are supposed to represent ores that have been regenerated from older sources with little or no change in isotopic composition. It is possible that the geochemical processes by which this regeneration and movement took place depleted the samples in silver content with respect to "normal" galenas.

A second correlation of trace element composition with isotopic constitution has been reported in the literature. The lead ores from Broken Hill,

N.S.W., Australia, described by Russell et al. (3) have lead isotope compositions that are related to the type of mineralization. The vein-type Thackaringa deposits are enriched in the radiogenic isotopes and are somewhat variable in isotopic composition compared to the conformable main lode and related ores in which the lead isotopes are homogeneously distributed. Qualitative spectrographic analyses of the Broken Hill region ores also revealed distinct differences between the Thackaringa and main lode types, particularly in the absence of detectable amounts of bismuth and tin in the former, compared to the presence of these elements in the latter. Other less striking differences were noted, but the silver content of the two groups was not significantly different. In view of the results of Cahen et al. (2) it was felt that the qualitative nature of the spectrographic analysis might be obscuring real differences in silver content.

TABLE I

RESULTS OF ANALYSES OF MAKE-UP SAMPLES

	Grams of lead	Micrograms of mercury taken	Micrograms of copper taken	Micrograms of silver	
				Taken	Found
Sample 1	0.5	30	30	30	30.6
2	0.5	30	100	30	29.1
3	0.5	100	30	30	29.8
4	0.5	30	100	15	15.6
5	1.0	20	25	45	43.5
6	0.5	30	50	20	20.0

To resolve this question, a quantitative chemical analysis procedure for the determination of trace amounts of silver in galenas has been developed and determinations made on the identical samples analyzed by Russell et al. (3). The procedure is one of double reversion employing dithizone as the reagent, and the details have been described fully by Bounsall and McBryde (4). Mercury and copper were interfering metals in the analysis. Results corrected for the blank, for the determination of silver in standard solutions of silver, mercury, copper and lead in proportions corresponding to the likely occurrence of these metals in galena, are given in Table 1.

Galena samples were dissolved in concentrated perchloric acid at a temperature of 250° to 275° F. This step replaced the more usual hydrochloric acid dissolution, since even small amounts of chloride ion interfere in the silver extraction by dithizone. Nitric acid was also unsatisfactory because of the large amounts of lead sulfate precipitated during the dissolving process. The lack of a trustworthy alternative procedure of analysis necessitated the use of a "spiking" technique to check the accuracy of the new procedure. Known amounts of silver were added to several of the ore samples immediately after dissolving, and subsequently subtracted from the value determined by analysis. The result obtained in this manner agreed well with "unspiked" determinations on other aliquots of the same ore samples.

The concentration of silver was expressed as p.p.m. lead rather than p.p.m. total sample since the lead sulfide purity of the samples was variable. Lead was determined for each silver analysis, essentially according to the chromate procedure outlined by Vogel (5). The results of the analyses, corrected for

TABLE II

Results of galena ore analyses		Average	Qualit. spec. results	
Sample numbers	P.P.M. silver in lead		Ag 3380.7	3382.9
Broken Hill Main Lode				
667	1,211, 1,241*, 1,197, 1,192	1,200	E	VS
668	427, 452, 422*	434	S	S
669	1,197, 1,246, 1,247*	1,230	S	S
670	572, 571, 578, 541*	566	VS	S
671	699, 734, 711*	715	S	S
672	993*, 999, 971	988	S	VS
Broken Hill type associated deposits				
912	1,780, 1,800, 1,810	1,800		
913	3,960, 3,950	3,960		
914	Single determination	2,670		
915	Single determination	699		
916	Single determination	459		
917	307, 293	300		
918	Single determination	1,920		
919	Single determination	529		
920	Single determination	1,190		
921	Single determination	1,300		
922	Single determination	704		
Thackaringa type deposits				
723	728, 716, 698	714	VS	VS
724	Single determination	721	—	—
725	437, 436, 443	439	S	S
726	852, 842, 858	851	S	S
911	Single determination	90		

E—extreme; VS—very strong; S—strong.

the blank and "spike" are shown in Table II. The samples marked with asterisks are those to which a "spike" equivalent to 10 micrograms of silver was added.

DISCUSSION OF RESULTS

The precision of the analyses is within 0.4 to 2.8 percent, based on the average deviation. The results indicate that there is no significant difference in the trace silver content between the Thackaringa type ores and the ores from

the Main Lode and associated deposits. From the geological location of the samples, (see Russell et al. (3)) there appears to be a variation with location, samples from or near the main lode at Broken Hill tending to have a higher trace silver content than samples from the outlying deposits. The most westerly sample (911) is quite low. One sample (668) taken from between the lenses in the main lode is distinctly lower in trace silver content than are the rest of the samples from the same lode. This apparent trend may not be real in view of the known tendency for silver to be concentrated in the near surface, oxidized or weathered zone of these ore deposits (Haddon King, personal communications). Much of the variation observed may be the result of this post-depositional enrichment process.

The analysis procedure used is the only chemical procedure described in the literature for the determination of trace amounts of silver in galena ores, and, as such, is a valuable and more accurate check on the results of analyses obtained by different means (i.e. spectrographic). For purposes of comparison, the original qualitative spectrographic analyses given by Russell et al. (1957) for silver are included in Table II. As can be seen, these analyses are insensitive to threefold variations in silver content and cannot therefore indicate trends in silver content over the concentration range determined. Although in this case trace silver analyses do not provide a means of distinguishing between lead ores that are isotopically different, the procedure would be applicable for determining the amount of silver in galena when the economics of the recovery of silver is the issue in question.

ACKNOWLEDGMENTS

The assistance of Dr. W. A. E. McBryde and Dr. A. D. Maynes in developing the chemical procedures used in this work is gratefully acknowledged. The author wishes to thank Eldorado Mining and Refining Company for the award of a scholarship which provided financial assistance for this work. The assistance of Dr. R. D. Russell and Dr. R. M. Farquhar in the geological interpretation of the results was quite valuable.

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CONTACT METASOMATIC COPPER MINERALIZATION AT THE CALZADA MINE, PERU

W. C. STOLL

The Calzada copper mine is 40 km S15°E of La Oroya, in the Cochas region of the Department of Junin, central Peru, and 20 km east of the crest of the Cordillera Occidental, at an altitude of 4,300 meters. It is reached through Llocllapampa on the Oroya-Huancayo highway, by way of the settlement of Pachicuyo. Chalcopyrite ore occurs in siliceous conglomerate beds, and was worked for a time by the Volcan Mines Co. of Lima. In 1956 about 700 tons of ore with an average grade of 5% Cu had been mined in an open cut from a conglomerate bed at least 6 m thick.

At the mine, a maximum stratigraphic thickness of 70 m of light colored conglomerates and quartzites, all more or less mineralized, is exposed, the strata striking N60°W and dipping 30°SW. The base is concealed by ground moraine. Overlying it with slight angular unconformity is a series, at least several hundred meters thick, of evenly stratified black, brown and dark red beds of which the lower members are graywacke. This formation resembles somewhat the continental Tertiary Rimac formation, exposed in the canyon of the upper Rimac river, near Casapalca. No intrusive rocks are exposed at the mine.

The conglomerates and interbedded quartzites are compact, cemented and metasomatized. They are little fractured and no mineralized faults were seen. The conglomerate consists of rounded, gray quartzite pebbles ½" to 2" in diameter and white chert pebbles, all in a matrix that has been almost totally replaced by silicates and sulfides. Microscopically the quartzite pebbles consist of equal-sized, rounded quartz grains (0.1 to 0.4 mm in diameter) and a few chert grains, with quartz cement and with small quantities of introduced sericite, calcite and chlorite. The chert pebbles, some of which show sedimentary bedding, are preferentially replaced by the epigenetic minerals. The chert appears as a micromosaic of quartz with some fine, probably sedimentary, material, and traversed by microveinlets of coarser, recrystallized quartz. Some of the matrix is quartz-cemented quartz grains. Most of it is composed of idiomorphic quartz and coarse white calcite, with smaller quantities of garnet, epidote, biotite, sericite, penninite, specularite and coarse pyrite and chalcopyrite. The pyrite is cubic; the chalcopyrite forms irregular blebs among the other metasomatic minerals and small veinlets in the chert pebbles. It was evidently the last mineral to crystallize.

Where not completely replaced, the quartzite layers interbedded with the conglomerates are, viewed microscopically, composed of a mosaic of rounded quartz grains 0.04 to 0.12 mm in diameter, cemented by quartz and replaced by spots of garnet with lesser quantities of biotite, epidote, sericite and calcite. One such bed has been converted almost totally to brown grossularite garnet occurring as tiny idiomorphic crystals showing twinning and growth zoning, together with traces of secondary quartz, sericite, chlorite and calcite.

The oligomict conglomerates and interbedded orthoquartzites are interpreted as second-cycle sediments, composed of detritus originating in some older series of quartzites and cherty limestones. They may be transgressive beach deposits, and may be found to be associated, at some horizon, with limestones, as is indicated by the large quantities of secondary calcite that replace them.

The overlying graywacke is composed microscopically of a fine dark paste in which are scattered angular fragments of quartz, chert and plagioclase, patches of fine-grained carbonate and pebbles and quartz grains derived from the conglomerates beneath. The graywacke has been altered apparently at the same time as the underlying formation. It contains micro-veinlets of epidote and spots of mixed diopside, garnet and epidote.

The suite of introduced minerals, rich in calcium, iron, copper and sulfur, is typically of contact-metasomatic origin. It is probably related to a concealed granitic intrusive, which has also affected, at greater depth, calcareous beds. The high degree of replacement of the dense siliceous sediments would indicate that the solutions possessed great permeative and reactive qualities.

The Norma deposit, near Lake Quiullupa and a few kilometers northwest of Calzada, is a similar occurrence, in the same conglomerate formation, which is here about vertical. The pyrite-chalcopyrite ore appears to follow a bed along which faulting has occurred in a direction N50°W, parallel to the bedding. The Manon mine, 10 km southwest of the Norma, contained similar replacement chalcopyrite ore, containing, in 1956, as much as 6% Cu, 0.5% Zn and 2 grams per ton each of gold and silver.

These occurrences of replacement ores of chalcopyrite might be akin to the contact-metasomatic ore deposits of southern Peru (Cuzco and Apurimac), such as Tintaya, Ferrobamba and others, which occur in limestone intruded by small masses of diorite or quartz monzonite. They are obviously distinct from the other types of copper occurrence in Peru. These are: 1) the polymetallic ores of the Cerro de Pasco type, generally with silver and minerals of arsenic and antimony, 2) the secondarily enriched porphyry copper ores of southwestern Peru (Toquepala, etc.), and 3) the red bed copper deposits of the Puno formation of southern Peru.

BUENOS AIRES, ARGENTINA,
January 18, 1961

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DISCUSSIONS

STRUCTURAL HISTORY OF THE BEAVERLODGE AREA

Sir: Mr. Chamberlain (3) apparently believes that criticism is best answered by suggesting that his paper has been misread when he states in answer to my discussion (6) "I am disheartened to see that Hill missed my remarks concerning. . . ." It is nevertheless still my contention that Chamberlain has not written a structural history of the Beaverlodge area.

The area included in Chamberlain's "Structural History" is approximately 500 square miles (2, Fig. 1). It is an area of great complexity with intrusions, gneisses, folds, cross-folds, schistosities, foliations, lineations, faults, joints, mineral deposits and sediments, of different types, origins, styles and ages.[1] To present, discuss, and interpret the history of this area Chamberlain uses six figures. Of these, five (2, Figs. 2–6) show only the ABC, St. Louis, or Black Bay faults, and apart from diagrammatic folding no cross-folds, lineations, foliations, schistosities, or joints are shown. Chamberlain's remaining figure (2, Fig. 1) covers 500 square miles, shows six additional faults, six generalized axial traces, and a random sprinkling of foliation and bedding symbols. The text itself contains little supporting or additional information. The six new faults are not mentioned. Schistosity "invariably parallels the bedding" (2, p. 480), cross-folding is "noted by Campbell . . . and the writer" (2, p. 480) while the main argument in support of a "primary bend" between the ABC and the St. Louis fault is "that the bedding, lineation and foliation of the Tazin rocks show little tendency to deviate from their regional northeast trend in the vicinity of the bend" (2, p. 483). As proof for this "little tendency to deviate" we are offered a map, scale $4\frac{1}{2}$ miles/inch with the nearest foliation symbols almost a mile from the bend (2, Fig. 1).

Chamberlain's evaluation and interpretation of the regional fault pattern is contained in two sentences. "Causes of large scale deflections of regional faults are poorly understood. They are probably controlled by the action of regional stresses operating in conjunction with heterogeneities of basement rocks or zones or weakness in the sub-crust" (2, p. 483). He does not discuss the problem further. He makes no attempt to explain what is meant by "localized, intense vertical forces" in his "tensional environment" (2, Table 1). The extent of his omissions may be gauged from Figure 1 (this paper), which shows the local faults. Chamberlain, as stated, discusses only three of these; presumably the others are discounted as being non-existent or irrelevant to a structural history or "typical of hundreds of minor breaks in the area" with "relative slip movements of generally less than 150 feet" (3).

[1] Criticism and suggestions by Michael Solomon are gratefully acknowledged.

Examples of the "minor breaks" omitted by Chamberlain are the North and Muskeg faults of Figure 2 (this paper). The dip of the North fault varies, on the surface, from 85 degrees north to 70 degrees south and the overall dip is taken to be essentially vertical. If Unit 3 is assumed to be a reliable marker—and it is the only unit of its kind in the area of Figure 2—movement along the fault may be calculated as (i) a right-handed strike-slip of 1,000 feet or (ii) with countryrock dipping eastward at 60 degrees, a

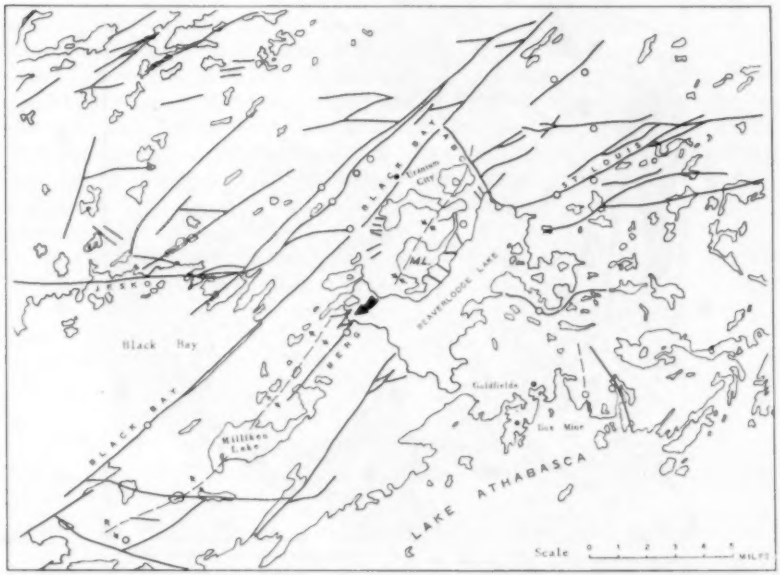

FIG. 1. The Beaverlodge Lake area, Saskatchewan, showing the prominent faults (solid lines), axial traces (dashes), and mineralized zones (circles). Martin Lake (M.L.). Based on published maps of the GSC and others. The area indicated by the large arrow is shown in Figure 2.

throw of 1,700 feet south side drop or (iii) obliquely from a combination of vertical and strike-slip movement. The Muskeg fault as determined from parallel and subsidiary (?) minor faults appears to dip vertically or steeply to the west. The apparent movement along the fault, as obtained by matching three contiguous units (Units 4, 5, 7a) at A with corresponding units at A′, is right-handed from 1,000 to 1,200 feet.

Chamberlain does not describe the types of cross-folds. There would appear to be several generations of "forces" (2, Table 1) and at least two periods of "northwest-southeast compression" the earlier followed by the development of cross-folds the "axial planes of which strike north to northwest" (2, p. 480) and for which Chamberlain postulates a "late change in

direction of force" (2, Table 1) a change presumably involving a swing in "force" through 75°. Is such a late change in direction of force necessary? How and when did the foliation develop? Have earlier foliations been folded and refolded?

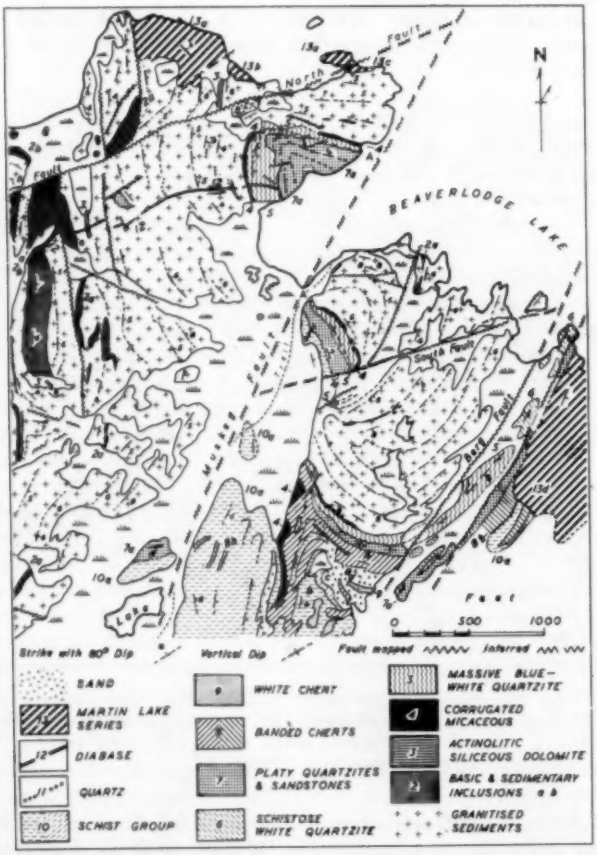

FIG. 2. Simplified geological map of the area indicated in Figure 1. Lineation, schistosity and other minor structures omitted. Strike lines are bedding-foliation. The *a, b, c, & d's* refer to rock types in the original paper (6).

Chamberlain appears to be unaware of other, local problems. Among others, he does not mention (i) the concentric structures of Martin Lake, (ii) the Milliken Lake antiform which heads towards the Martin Lake syncline, (iii) the intrusive (?) and non-intrusive (?) quartzites,[2] (iv) the influence

[2] The two types of quartzite are visible at the Box mine of Figure 1 (Hill 1952 unpub. report and Hill (5, p. 229)).

of graphitic schists on local tectonics, (v) the peculiar elongated (in *b* ?) cobbles of quartzite near some local faults, (vi) the environment and provenance of the so-called Athabascan.

In discussing the Athabascan Chamberlain "accords with the view" (2, p. 480) that the Athabascan series south of Lake Athabasca is "equivalent to the Athabascan series in the Beaverlodge area." Gussow, however, in a paper published 4½ months previously [3] (4) showed that the two series are not equivalent and that the beds in the Beaverlodge area, renamed the Martin Lake series, are probably much older than the Late Proterozoic to which they had hitherto been assigned. Gussow further suggested that whereas the true Athabascan is marine and Palaeozoic the Martin Lake series are of freshwater origin and late Early Precambrian. As Chamberlain uses the Martin Lake beds to determine and date the movement along his three faults and uses the lack of folding in the Athabascan south of Lake Athabasca (2, p. 493) as evidence that his second "horizontal compression exerted northwest-southeast" was "more localized and less intense" than his first, the reader might be excused for thinking that this new information might have been added as a footnote to the page proof.

In conclusion, the structural history of the Beaverlodge area as presented by Chamberlain is inadequate.

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INTRIGUING EXAMPLES OF GEOLOGY APPLIED TO INDUSTRIAL MINERALS

Sir: In Vol. 55, page 635 of this Journal, Dr. Joseph L. Gillson stated that "in general, fluorspar is a lone wolf, occurring in deposits that are away from mining districts noted for the occurrence and production of other commercial minerals."

[3] Chamberlain's paper was received by the editor six months prior to the appearance of Gussow's paper.—*Editor.*

Although Dr. Gillson went on to give a few exceptions to the above statement he did not go nearly far enough in giving exceptions. If he had gone far enough and included low-grade epigenetic fluorspar deposits the exception would, I believe, become the rule. His statement might then have been, "in general, fluorspar is closely associated with one or more of the following metals—tin, tungsten, beryllium, molybdenum, copper, zinc, lead, silver, gold, uranium and manganese, and is in districts noted for the production of one or more of said metals. To date, however, it has been possible to obtain much of the required fluorspar from the less common, sometimes isolated, high-grade fluorspar deposits that do not appear to have the normal close association with the above metals or at least are not yet noted for production of one or more of the metals."

A few of the many examples of the association of fluorspar with some of the above metals are found in such districts as North Pennine, England; Parral, Mexico; Illinois-Kentucky Fluorspar District; Climax, Cripple Creek and Jamestown, Colorado; Hansonburg, New Mexico; Mt. Wheeler, Nevada; Star and Thomas Range, Utah; and Bayhorse, Idaho.

The thought-provoking work of Dr. Peters, referred to by Dr. Gillson, recognizes that fluorspar is a relatively common minor gangue mineral in nearly all types of ore deposits, but that it is of commercial importance mainly in shallow-seated deposits. With depletion of the high-grade fluorspar deposits, increased demand for fluorspar and improved extractive metallurgy the more complex metal-bearing fluorspar deposits will become more and more important as a source of fluorspar. As that happens the lone wolf, shallow-seated label for fluorspar will be recognized as misleading.

Silver has long been recognized as playing an important role in the discovery of and exploitation of many important mining districts. Near-surface high-grade silver ore in many places led to discovery of complex base metal ores at depth where the silver grade declines but still helps pay the cost of efficient exploration, development, and mining.

Epigenetic fluorspar and manganese can play roles similar to silver in leading to the discovery of new centers of mineralization. They can also be helpful in converting small districts into major districts and revival of ghost camps. Neither of these gangue minerals will have the glamour of silver and rarely will they alone be capable of leading to important discoveries. They will, in most cases, merely pay part of the cost of exploration, development, and exploitation. Some other associated product such as silver, lead, copper or beryllium will need pay its share of expenses.

This is written to help erase the misleading lone wolf label that Dr. Gillson would attach to fluorspar, for I believe that such misplaced labels interfere with the efficient exploration, development, and exploitation of mineral resources. This is likely to be particularly true at a time when national minerals policies are being given much thought.

Dooley P. Wheeler, Jr.

Challis, Idaho,
Jan. 19, 1961

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REVIEWS

Geology of Ore Deposits (Geologiya Rudnykh Mestorozhdenii). Academy of Sciences of the USSR. May–June, 1959; Vol. 1, No. 3 (in Russian).[1]

E. I. SHATALOV, *Principles of outlining the metallogenetic regions* (Pp. 3–33).

According to the general trend to advance applied science that has an immediate bearing on the industrial power of the country, the author emphasizes metallogeny as the most important branch of geology.

After reviewing the terminology and definitions used by the students of metallogeny, a modified classification of metallogenetic areas is suggested:

Size of ore-bearing areas	Form of ore-bearing areas	
	Elongated, linear	Non-linear
Very large	Metallogenetic belt	Metallogenetic province
Large	Metallogenetic zone	Metallogenetic region
Medium	Ore-bearing zone	Ore-bearing district Ore-bearing center
Small	Ore-bearing field	

A metallogenetic belt is a folded structure formed in place of a geosyncline during a major orogenic cycle. It is the result of tectonic and magmatic processes with affiliated mineralization. The belts, such as the Pacific, Mediterranean, and Uralian, are up to 10,000 km long and include the metallogenetic provinces, zones, regions, and smaller units. A metallogenetic province, such as the provinces of the eastern Transbaikal region and parts of the Pacific metallogenetic belt, is characterized by a complicated structure, formed in place of a geosynclinal region, within the limits of a shield or platform, during two or more orogenic epochs. Maps of metallogenetic belts and provinces on a scale not less than 1 : 1,000,000 are used to outline the potential ore-bearing zones, regions and districts.

A metallogenetic zone is an elongated ore-bearing structure, which is up to 1,500 km long, and formed during a definite stage of formation of folded belts and platforms with typical mineralization, depending on the character of sedimentation, tectonic regime, and magmatic processes. These conditions result in ore formations characterized by specific associations of major and minor metals. The term ore-bearing belt is synonymous with the metallogenetic zone. The metallogenetic regions are analogous to zones. Maps of zones and regions on scales of 1 : 1,000,000 to 1 : 200,000 serve to direct the geological prospecting, exploration, and detailed survey on scales of 1 : 50,000 to 1 : 10,000.

An ore-bearing district is an area 100 to 250 km long and up to 150 km wide, characterized by specific geological features, ore formations, or types of economic mineral deposits. The ore-bearing zone is 40 to 50 km long and several kilometers wide. The ore-bearing center is of irregular outline and consists of a group of ore-bearing fields or separate ore deposits with similar mineralization, affiliated with an intrusive body. The ore-bearing center is generally located at an inter-

[1] Regular preparation of reviews of this journal is supported by a grant from the National Science Foundation.

section of folds with a system of faults. An ore-bearing field occupies an area of 10 to 20 square km and consists of a group of genetically related deposits. Maps of ore-bearing fields are used to locate the prospective deposits, especially the blind ore bodies.

A. I. GINZBURG, *New types of rare metals deposits* (Pp. 34–47).

Besides the littoral deposits of heavy minerals of recent origin, placer deposits were discovered in Tertiary, Cretaceous, Devonian, and Proterozoic formations. The younger placer deposits consist predominantly of ilmenite which is replaced by leucoxene (commonly containing columbium) and hematite-rutile aggregate in the older, metamorphosed placers. The metamorphosed placers are generally characterized by hydrothermal uranium and thorium mineralization. In this case uranium and thorium were derived by hydrothermal solutions from metamorphosed placers containing cyrtolite, monazite, brannerite, and some uranium-bearing tantalates and columbites.

Deposits of halogens, coal, bituminous shales, phosphates, some bauxite, and sedimentary iron and manganese ores also contain recoverable rare elements.

All known hypogene deposits of lithium, beryllium, cesium, columbium, tantalum, zirconium, and rare earths are genetically related to granitic and alkaline intrusive complexes. The magmatic and pneumatolytic-hydrothermal deposits are considered as more promising than the deposits of rare elements of pegmatitic type. Pegmatites containing lithium are an exception. Besides placers of different age, the most prospective economic deposits of rare elements are: (1) the stratified ultrabasic intrusions with loparite enriched horizons (cerium-columbium deposits); (2) surface regoliths containing monazite, columbite, and fergusonite; (3) spodumene-bearing pegmatites (tantalum, beryllium, and lithium deposits); (4) granitic massifs with the enclosing altered rock formations (beryllium-bearing greisenized granites, stockworks, and veins); (5) beryllium-bearing skarns and metasomatic fluorite deposits with chrysoberyl, phenacite, and helvine; (6) micatized pneumatolytic-hydrothermal formations in serpentinites and chlorite-talc deposits containing beryllium; (7) albitized nepheline syenite and the enclosing altered sedimentary rocks containing pyrochlore, zircon, and thorium; (8) albitites with zirconium-columbium-ittrium mineralization; (9) albitized granitoid massifs with columbite and zircon; (10) pyrochlore carbonatites (columbium deposits); (11) hydrothermal deposits of rare earths formation (cerium and thorium-cerium deposits).

Each type of rare metals deposit is described from the petrological and structural points of view, together with minerals inherent to each genetic type of deposit.

L. N. OVCHINNIKOV AND V. G. MAKSENKOV, *The geological structure of the Gora Vysokaya ore-bearing field* (Pp. 48–61, map).

The deposit of Gora Vysokaya is located in the Tagil region of the Urals where about 20 contact-metasomatic iron ore deposits occur in an area of 80 square km. The possibility of discovering new deposits in this area is still good. All deposits of the Gora-Vysokaya area are genetically and structurally associated with the synenite massif which is 340 to 360 million years old and is located in a broad horizontal flexure. The country rocks are of Upper Silurian age, dipping 50 to 70° to the east, and consist of limestones intercalating with tuffs. Rocks of syenite complex demonstrate an increase in basicity toward the contact with adjacent gabbro-norites. The syenite massif is surrounded by a zone of contact metamorphic alteration which is one km wide. The mineralization was preceded by potassium feldspathization. The age of iron mineralization is close to that of

syenite. The ore bodies are structurally associated mainly with tension fractures, open fractures parallel to the bedding, and less with tension fractures in gabbro bordering the flexure, and with fractures of shearing type.

An analogy is drawn between the Gora Vysokaya type of mineralization and the iron deposits of Gornaya Shoriya in Siberia. The authors suggest that during exploration more attention should be paid to flexures as major ore-controlling structures.

E. M. NEKRASOV, *Structural peculiarities of the lead-zinc deposit of Zambarak in eastern Karamazar* (Pp. 62–73).

The deposit is located on the southern slopes of the Kuraminsky Range in Uzbekistan. The country rocks consist of Upper Carboniferous effusives which form a large syncline. The ore bodies are located in the axial part of this syncline and are represented by galena-barite and quartz-hematite veins, stockwork zones, and disseminated types of mineralization. A structural study of veins indicates that the deposition of ore minerals was localized along the shear fractures and sometimes along the tension fractures. The formation of these fractures was evidently simultaneous with the process of mineralization. Fractures striking 23–33° and 45–60° to the northeast are characterized by high-grade mineralization. Low-grade lead and zinc ore formed in fractures containing gouge, which impaired the circulation of mineralizing solutions.

D. I. GORZHEVSKY, *Some features of distribution of belts of tin and lead-zinc deposits* (Pp. 74–80).

The author analyzes the tectonic features of distribution of the ore-bearing zones in the Rudnyi Altai and eastern Transbaikal regions. In Rudnyi Altai the sequence is 10 to 12 km thick and consists of Middle and Upper Devonian and Lower Carboniferous volcanic rocks, intercalating with terrigenous deposits and limestones. The lead-zinc deposits are structurally associated with a complex of minor intrusions. To the southwest of Rudnyi Altai is located the Kalba synclinorium containing Silurian marine deposits. The main ore deposits within this synclinorium are represented by rare earth pegmatites and tin-tungsten mineralization, genetically affiliated with the Upper Paleozoic granite intrusions. It is concluded that the lead-zinc mineralization is associated with the Hercynian geosyncline, which is superimposed on the Caledonian anticlinal uplift, while the tin-bearing zone of Kalba is located within the boundaries of the Hercynian geosyncline that inherited the Lower Paleozoic geosynclinal subsidence.

In the northwest of the eastern Transbaikal region there is a belt of gold-molybdenum deposits. The central part of the region is occupied by tin-tungsten deposits, and the southeast is characterized by the lead-zinc deposits. The tin-tungsten deposits are of Upper Paleozoic and mainly of Upper Jurassic age, and are affiliated with the central part of the relict Mesozoic geosyncline, which inherited the Paleozoic synclinal structure and contains 12 km of sediments. On the other hand, the lead-zinc mineralization developed in the transitional marginal zones of the Mesozoic geosynclines with 6 km of Paleozoic sediments superimposed over the Paleozoic geanticlinal uplift. The deposits of cassiterite-sulfide formation are an exception, and are localized in the area of lead-zinc mineralization.

Beginning with the Paleozoic, there were two structural zones with different rates of subsidence. The process of ore formation is closely related to the process of evolution of the earth's crust, involving the accumulation of sediments, tectonic movements, associated magmatic activity, and metamorphism.

A. A. Filimonova, *Textures due to unmixing of solid solutions in ores altered by metamorphism* (Pp. 81–88, 2 plates).

A comparison of natural unmixing textures with textures produced experimentally is a means of reconstructing the temperature conditions of mineralization. The pattern of unmixing textures formed by diffusion in solid state reflects not only the process of cooling but also the primary texture of mineral aggregates.

An experimental heating of pyrite-bornite ore at 360°C for 15 minutes results in formation of reaction rims in areas adjacent to pyrite. The variety of unmixing textures in small areas is explained by the character of primary intergrowths. Heating of mineral aggregates results in regrouping of products of unmixing and in growth of inclusions on account of reduction of their number. Compounds of intermediary character form between two reacting minerals. Pores are formed due to the escape of sulfur in areas of chalcopyrite-bornite intergrowths when ore is heated under atmospheric conditions.

Ore samples from the Kabanskoe deposit in the Urals have textures similar to those obtained by experimental means. Reaction rims of chalcopyrite around pyrite grains and pores are interpreted as the result of irregular heating caused by friction during dynamic metamorphism.

An example of thermal metamorphism is illustrated by the unmixing textures in bornite-pyrite ore from the Levikha deposit in the Urals. In this deposit the unmixing disappears within a distance of 2.5 m from a 9 m thick dike. At the contact the unmixing results in formation of chalcopyrite lattice texture. Away from the dike the ore is represented by bornite with pyrite only. The thermally altered ore is porous because of the removal of material during metamorphism.

Z. P. Potapova, *The Pobednoe iron deposit* (Pp. 89–102).

The deposit is located in the Kolyma River basin in northeastern Siberia and is an unusual example of association of sedimentary hematite ore with dolomite. The series containing iron ores is presumably of Lower Devonian age and consists almost entirely of dolomitic rocks. The upper part of the series is represented by the subseries containing the iron ore bed. The subseries consists of hematite- and magnetite-bearing carbonate rocks (dolomites and, to some extent, ankerites). The ore bed is generally 3–6 m thick and attains in places a thickness of 20 m. The known extension of the ore bed is 18 km. Iron ore is represented by a microcrystalline aggregate of hematite, which constitutes 89–96 per cent of the ore.

During the Upper Silurian the region was represented by a marine basin of normal salinity. During Lower Devonian time the basin became shallow and a lagoon formed, creating saline conditions unfavorable for marine life. The accumulation of dolomites followed by the deposition of iron ore indicates the beginning of transgression over the peneplanized lateritic land surface and the change of the arid climatic conditions to humid conditions. These conditions are reflected in the corresponding sedimentary facies. Iron was removed from the lateritic soil in colloidal or bicarbonate solutions. The precipitation of iron in the form of cryptocrystalline hematite followed after these solutions encountered the alkaline water of the marine basin. Ankerite and magnetite formed within the sediment under slightly reducing conditions.

L. M. Lebedev, *The morphologic features of quartz and cassiterite in the Shakh Shagaily deposit* (Pp. 103–114, 8 plates).

The deposit is located in the Karaganda region of Kazakhstan and represents a transition between the deposits of cassiterite-quartz and cassiterite-sulfide forma-

tions. The deposit has no economic value but is interesting from the genetic point of view since it contains metacolloidal aggregates of cassiterite and quartz.

The deposit is associated with the fractured greisenized zone within leucocratic granites. The quartz-cassiterite aggregates form cement between the fragments of greisen. The fine-grained and the colloform aggregates of quartz and the metacolloidal aggregates of cassiterite form the middle zone of geodes. The central parts of the geodes consist of drusy quartz. The author assumes that flakes of colloidal SnO_2 coalesced and formed dense spherical bodies. This process was probably similar to the formation of spherulites of sphalerite during an experimental coagulation of colloidal zinc solution.

It is concluded that the colloidal solutions played a major role in the process of mineralization. After deposition the quartz-cassiterite gel was the subject of alteration which produced cryptochrystalline aggregates of metacolloidal quartz. The coarse crystalline aggregates formed later during the cumulative recrystallization. The metacolloidal cassiterite recrystallizes only in areas of the most intensive process of recrystallization.

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The Geological Evolution of North America—A Regional Approach to Historical Geology. By THOMAS H. CLARK and COLIN W. STEARN. Pp. 434. Ronald Press Co., New York, 1960. Price, $7.50.

This large volume is designed for beginning courses in Historical Geology, and is stated to offer a fresh approach to the historical geology of North America. It is organized as a series of regional studies to provide continuity in tracing the geologic history of the continent, rather than by a chronologic treatment by geologic periods. The authors consider the historical development of the continent in terms of its three major units—bordering synclines, the stable interior, and the Canadian Shield. After the introductory chapters, sedimentation is given first consideration.

There are five parts and an appendix giving a summary of the plant and animal kingdom, and a simplified classification of them. Part II takes up the Appalachian Geosyncline—its early evolution, orogenic climax, Ouachita Geosyncline, and the Appalachian Mountains and the coastal plains. Part III—the Cordilleran Geosyncline—treats of the beginnings and Paleozoic history, Mesezoic and Cenezoic orogeny and sedimentation. Part IV—the Continental Nucleus and Northland—is divided into the Arctic Islands, the Canadian Shield, and the Great Ice Age.

Part V is the Evolution of Life, divided into Precambrian and early Paleozoic, late Paleozoic, Mesozoic, and Cenozoic.

The book is an example of fine writing. It is a large page with double colums and is excellently illustrated and printed. It should serve as an unusually interesting textbook for a beginning course.

Paleogologic Maps. By A. I. LEVORSEN. Pp. 174; black and white maps. W. H. Freeman & Co., San Francisco, 1960. Price, $6.00.

This volume by an outstanding petroleum geologist discusses the preparation of paleogeologic maps and the interpretations that can be made from them. Some of the problems to which such maps are applied are unconformities, pre-unconformity structures, shore line truncation, time of deformation, location of wedge belts of

sediments, time of oil and gas migration, and earth ruptures. They are essential for the interpretation of Levorsen's "layer cake" geology. The examples of maps given are chosen because of some special element or interpretation, or from which deductions may be drawn.

The first chapter deals with sources of data, and preparation and interpretation. The next chapter takes up the uses of paleogeologic maps as applied to the interpretation of structure and geologic history. A third chapter draws examples from a number of oil fields of the United States and one each from Russia and Venezuela. A fourth chapter gives consideration to regional areas, and lists 13 regions in the United States and other oil-producing countries. The last chapter, "Continental Areas" considers North America, Siberia, Continental Drift, and the World References, and selected readings complete the book.

This book is full of meat, as are all the writings of Dr. Levorsen. It should be useful to students of stratigraphy and petroleum geology, as well as to practicing geologists.

From Theory to Practice in Soil Mechanics. (Selections from writings of Karl Terzaghi.) Pp. 425; figs. John Wiley & Sons, New York, 1960.

This large volume was prepared by L. Bjerrum, A. Casagrande, R. B. Peck, and A. W. Skempton, who also contribute chapters on his life and achievements. The selections, given in his own words, supplemented by explanatory comments, show how Terzaghi proceeded, step by step, to establish the fundamental principles of soil mechanics and then to use them as tools in engineering practice. Thus there is provided demonstrations, especially to young engineers, the techniques for successfully practicing soil mechanics.

Of the five parts, the first covers his life and achievements, the second deals with the aim, scope, and methods of soil mechanics. The next two parts are writings of varied subjects taken from selected professional reports dealing with soil mechanics and problems of engineering geology. Part five is a bibliography.

The material is carefully selected and nicely printed and illustrated. The book should prove to be a good reference work for those interested in soil mechanics and geological engineering.

Tropical Africa. By GEORGE H. T. KIMBLE. 2 Vols.; pp. 1110; figs. 47; tables 66. Twentieth Century Fund, New York, 1960. Price, $15.00.

These volumes cover the vast area of Africa south of the Sahara and north of the Union of South Africa—an area very much in the news of late. This study was authorized by the Twentieth Century Fund, a non-profit philanthropic foundation, and drew on forty-six experts and specialists in its preparation, which took seven years.

Volume I, *Land and Livelihood*, covers such subjects as early economic life, population, physical features, farming, forestry, food, minerals, land settlement, industrial progress, merchandising, and the workers. Volume II, *Society and Polity*, considers the old and the new order, social changes, problems of ignorance and sickness, colonial policies, nationalism, government, growth, and the new elite.

The two volumes constitute an encyclopedic survey of this interesting and puzzling territory and provide knowledge for guidance in the struggles of the peoples of Africa to reach freedom and independence. This scholarly work, in incisive style should be read by all interested in Africa.

Introduction to Geophysical Prospecting, 2nd Edit. By MILTON B. DOBRIN. Pp. 446. McGraw-Hill Book Co., New York, 1960. Price, $9.50.

This revised edition brings up-to-date the considerable new information regarding geophysical prospecting that has appeared in the nearly ten years since the previous edition was written. Much of the book has been rewritten and other parts thoroughly revised. The book covers all branches of geophysical prospecting, both petroleum and mining, although petroleum prospecting is naturally given more attention than that for mining since it more widely undertaken and the author's experience has been largely in petroleum. Emphasis is given on results of actual field surveys.

After an introductory chapter, seismic principles, instruments, methods and interpretation are treated in the next six chapters. This is followed by gravity prospecting in five chapters dealing with the principles, instruments, field measurements and reduction, and interpretation. Magnetic prospecting is dealt with in four chapters, followed by a chapter each on electrical and radioactive prospecting. A final chapter takes up the integration of geophysical methods as illustrated by case histories.

The book contains a minimum of mathematics and is meant for students of geology and mining who plan to enter the fields of petroleum and mineral exploration, and for professional geologists and geophysicists. It should prove to be a valuable text and reference book.

BOOKS RECEIVED

ROGER L. AMES AND JOHN E. COTTON

Recent Sediments, Northwest Gulf of Mexico—A Symposium. Edited by F. P. SHEPARD et al. Pp. 394. American Association Petroleum Geologists, Tulsa, Oklahoma, 1960. *A symposium of results of Project 51 of American Petroleum Institute. 14 papers dealing with various phases of this subject.*

Major Activities in the Atomic Energy Programs January–December 1960. Pp. 544. Price, $1.75.

Annual Report of the Director of the Geophysical Laboratory 1959–1960. PHILIP H. ABELSON. Pp. 43–187; pls. 4; figs. 77. Carnegie Institution of Washington, Geophysical Laboratory Paper No. 1340, Washington, D. C., 1960. *A summary of investigations from July 1, 1959–June 30, 1960.*

Historical Statistics of Minerals in the United States. SAM H. SCHURR and ELIZABETH K. VOGELY. Pp. 42; tbls. 274. Price, $1.00. Resources for the Future, Inc., Washington, D. C., 1960. *A reprint of the Minerals Chapter of "Historical Statistics of the United States—Colonial Times to 1957."*

Introduction to Geology (An Outline). Revised Edition. ALFRED LIVINGSTON, JR. Pp. 55; figs. 15. N-P Publications, Palo Alto, California, 1960. *An outline prepared to present fundamental principles of geology without using the aid of examples.*

Cronica Bibliografica. Pp. 48. Publicaciones J-1958, Instituto de Fisiografia y Geologia, Rosario, Republica Argentina, 1959. *Annotated bibliography of notable foreign articles translated into Spanish.*

Annual Report of the Geological Survey Department, 1959. Pp. 39; tbl. 1. Map, scale 1/2,000,000. Price, 6/6. Bechuanaland Protectorate, Lobatsi, 1960. *Asbestos production dropped following a drop in price of fiber early in 1959.*

Geology of the Rocher Deboule Range. A. SUTHERLAND BROWN. Pp. 78; pls. 14; figs. 15; tbls. 2. British Columbia Dept. of Mines and Petroleum Resources Bulletin 42, 1960. *Valuable ores of tungsten and copper are present.*

Report on the Geological Survey Department for the Year 1959. Pp. 67. British Guiana, Georgetown, 1960. *This annual report contains a summary of important mineral products and a series of preliminary report abstracts by field geologists.*

Geology of the Otay Bentonite Deposit, San Diego County, California. G. B. CLEVELAND. Pp. 16; pl. 1; figs. 5; tbls. 3; photos 6. Price, $1.00. California Division of Mines Special Report 64, San Francisco, 1960. *Six or more beds of bentonite range in thickness from less than a foot to 4 feet.*

Maps of the Equatorial African States—Congo, Gabon and Tchad. All maps scale 1/500,000. Explanatory text in French: **Quanda-Djalle-Est, Quanda-Djalle-Ouest, Sibita-Ouest, Fort Archambault-Est, Adré, Libreville-Est, Fort-Crampel-Ouest, Mayoumba-Est.** Institut Equatorial de Recherches et d'Etudes Géologiques et Minières, Brazzaville, 1958–1960.

Report of the Director of Geological Survey for the period 1st April, 1958 to 31st March, 1959. Pp. 23. Price, 3s. Ghana Geological Survey, Accra, 1960. *Hematite ores near Shrene in the northern region of Ghana range from about 35 percent to 50 percent Fe. Some 100 million tons of ore are available for open-cast working.*

Study of Two Idaho Thorite Deposits. JOSEPH NEWTON, DENIS LEMOINE, E. N. ADAMS and others. Pp. 53; figs. 12; tbls. 11. Idaho Bureau of Mines and Geology Pamphlet 122, Moscow, 1960. *The results of a geologic and metallurgical study of thorite in relation to its possible economic source of thorium.*

Differentiation of Caseyville (Pennsylvanian) and Chester (Mississippian) Sediments in the Illinois Basin. ELWOOD ATHERTON, G. H. EMRICH, H. D. GLASS, and others. Pp. 36; figs. 14; tbls. 3. Illinois Geological Survey Circular 306, Urbana, 1960. *Regional variations in both Caseyville and Chester sediments render recognition difficult in many areas.*

Seventh Biennial Report of the Division of Geological Survey and water Resources of the Department of Business and Administration State of Missouri . . . , July 1, 1958–June 30, 1960. T. R. BEVERIDGE. Pp. 57; figs. 3. Rolla, 1960. *Work is under way on a new geological map of the state. Low wage scales for Survey geologists hampers efficient activity.*

Fracture Systems and Tectonic Elements of the Colorado Plateau. V. C. KELLEY and N. JAMES CLINTON. Pp. 104; pls. 25; figs. 9. Price, $2.75. University of New Mexico Publications in Geology Number 6, Albuquerque, 1960. *The authors conclude that the fractures are of many ages and of several different origins.*

Coal Resources of the Upper Part of the Monongahela Formation and the Dunkard Group in Ohio. G. H. DENTON. Pp. 50; figs. 15; tbls. 16. Ohio Division of Geological Survey Report of Investigations No. 38, Columbus, 1960. *An estimated 3.9 billion short tons of coal occur in the Monongahela and Dunkard beds.*

Mineralogical Characteristics of the Oxidation Zone of Silesia—Cracow Zinc and Lead Deposits. WITHOLD ŻABIŃSKI. Geol. Trans. No. 1. Pp. 99; figs. 22; pls. 7. (12-page digest in English.) Polska Akademia Nauk, Warsaw, 1960. *Description of minerals; limonites; general character of oxidation zone.*

The Dundas Group in the Queenstown Area. M. SOLOMON. Pp. 33–50; pl. 1; figs. 4; tbls. 2. University of Tasmania Dept. of Geology Publ. 82, Hobart, 1960.

Texas Fossils, An Amateur Collector's Handbook. W. H. MATTHEWS, III. Pp. 123; pls. 49; figs. 26. Price, $1.00. Bureau of Economic Geology, The University of Texas, Guidebook 2, Austin, 1960. *A booklet on where and how to collect fossils and how to prepare identify, and catalog specimens.*

Radioactive Mineral Deposits of Wyoming. W. H. WILSON. Pp. 41; fig. 1. Wyoming Geological Survey Report of Investigations 7, Laramie, 1960. *Most of the current uranium production comes from deposits in the Eocene Wind River formation in the Gar Hills area.*

University of California Publications in Geological Sciences—Berkeley and Los Angeles, 1960.

Vol. 34, No. 6. Displaced Miocene Molluscan Provinces Along the San Andreas Fault, California. C. A. HALL, JR. Pp. 281–301; tbls. 8; chart 1; maps 3. Price, $1.25. *Right-lateral slip displacement of approximately 120 miles is indicated in the method used.*

Vol. 38, No. 2. Mississippian and Devonian Paleontology and Stratigraphy Quartz Spring Area, Inyo County, California. R. L. LANGENHEIM, JR., and H. TISCHLER. Pp. 89–150; pls. 14–15; figs. 18. Price, $1.25.

Vol. 39, No. 1. Late Pliocene Floras East of the Sierra Nevada. D. I. AXELROD and W. S. TING. Pp. 118; pls. 24; figs. 10; tbls. 2. Price, $2.50.

Geological Survey of Canada—Ottawa, 1960–1961.

Bull. 61. Petrology of the Gneisses of Cumberland Sound, Baffin Island, Northwest Territories. G. C. RILEY. Pp. 68; pls. 10; fig. 1. Map, scale 1/506,880. Price, $1.00. *Precambrian metamorphic rocks are assigned to five metamorphic zones.*

Paper 60-29. Ground-Water Resources of Sumas, Chilliwhack, and Kent Municipalities, British Columbia. E. C. HALSTEAD. Pp. 39; fig. 1; tbls. 8. Maps, scale 1/63,360. Price, 50 cents. *Development of available ground-water resources should provide adequate supplies.*

Paper 60-30. Iron-Formations and the Labrador Geosyncline. G. A. GROSS. Pp. 7. Map, scale 1/1,013,760. Price, 50 cents. *The Labrador geosyncline is divided into 3 geological divisions for economic considerations of iron-formations.*

Paper 61-1. Summary Account of Carboniferous and Permian Formations, Southwestern District of MacKenzie. P. HARKER. Pp. 9. Price, 25 cents. *Five new formation names are proposed.*

Aeromagnetic maps 774G, 789G, 793–796G, scale 1/63,360—**Sambro, Halifax, Oxford, Wolfville, Parrsboro, and Springfield, Nova Scotia.**

Aeromagnetic maps, scale 1/63,360, **950G–954G: Thunder Bay District Ontario: 955G–959G: Kenora District, Ontario; Geology map 25-1960,** scale 1/253,440, **Glenlyon, Yukon Territory; Geology map 27-1960,** scale 1/12,000, **Chisel Lake, Manitoba.**

Aeromagnetic maps, scale 1/126,720, **1015G–1024G, Gulf of St. Lawrence; 940G–949G,** scale 1/63,360, **Goldsborough, Whitewater, Greenmantle, Pruner Crerar, Jervis Bay, Tologan, Wigwascence, Michikenis, and Wunnummin Lakes, Ontario.**

Aeromagnetic maps, scale 1/63,360, **Kenora District Ontario maps 850G–859G, 870–889G, 900–909G, 922–929G, 934–939G, 945–949G; Thunder Bay District,**

Ontario, 920–921G, 930–933G, 940–944G; Nova Scotia and Prince Edward Id., maps 230, 240, 818, 826–849, 1010–1014, 1025–1026; New Brunswick, 815–817, 824; Index to map sheets, Ontario No. 7; Aeromagnetic Survey across Cordillera, map 749G, scale 1/253,440.

Aeromagnetic maps, scale 1/63,360, **Thunder Bay, 960–962G, 970–971G; Thunder Bay and Kenora, 963–964G, 872–973G; Kenora District, 965–969G, 974–979G, Ontario; Colored geology, Cracking Stone, Saskatchewan 1095A and Trout River, Newfoundland, 1086A; Surficial geology, Courtenay, Vancouver Island, B. C., 32-1960.**

Aeromagnetic maps, scale 1/63,360, **Ontario, Thunder Bay and Cochrane, 990–994G; Kenora, 995–999G, 1006–1009G; Cochrane, 1000–1003G; Cochrane and Kenora, 1004–1005G; surficial geology, 36–37, 1960, Mt. Steward and Souris, Prince Edward Island; maps 26 and 35, Island Lake, Ontario and King's Point, Newfoundland.**

Aeromagnetic maps, scale 1/63,360, **Ontario, Kenora District, 860–869G, 984–988G; Thunder Bay District, 980–982G; Kenora and Thunder Bay, 982–983G.**

Geological Survey of Connecticut—Middletown, 1960.

Quadrangle Rept. 10. The Surficial Geology of the Wallingford Quadrangle. S. C. PORTER. Pp. 42; figs. 18. Map, scale 1/24,000. *A large quantity of sand and gravel is available for mining.*

Quadrangle Rept. 9. The Bedrock Geology of the Naugatuck Quadrangle. M. H. CARR. Pp. 25; figs. 5; tbls. 2. Map, scale 1/24,000.

Geological Survey of Japan—Hisamoto-chō, Kawasaki-shi, 1960.

Bulletin of the Geological Survey, Vol. 11, No. 6. Pp. 70. *Four articles on geophysical prospecting for uranium deposits.*

Geologic maps: Akita, scale 1/500,000; **Gohyakkoku, Harutachi, Yagishirito,** scale 1/50,000; **Mineral Province of Japan: Mineralization of Neogene Period, Mineralization of Mesozoic to Paleogene Period, Mineralization of Paleozoic Period,** all scale 1/2,000,000; **Akita and Ogashima and Mito,** scale 1/200,000; **geologic maps of the coal fields of Japan with cross sections, Hokusho Coal Field (Kambayashi, Utagaura, Senryu, Saza, Kusudomari, Yamanoda),** scale 1/10,000; **General Geologic Map of Kokusho Coal Field with accompanying map of geologic profiles,** scale 1/25,000; **Geologic Map of Hokusho Coal Field at Sea Level.**

Kansas Geological Survey—Lawrence, 1960.

Bull. 148. Geology and Ground-Water Resources of Douglas County, Kansas. H. G. O'CONNOR. Pp. 199; pls. 9; figs. 10; tbls. 8. *Data regarding present and potential ground-water developments, together with the effects of pumping on discharge, recharge, and quality of the water, are summarized for the principal aquifers.*

Oil and Gas Investigations 22. Preliminary Regional Structural Contour Map on Top of Mississippian Rocks in Kansas. D. F. MERRIAM. Map, scale 1″/10 miles.

Oil and Gas Investigations 23. Preliminary Regional Structural Contour Map on Top of "Hunton" (Silurian-Devonian) Rocks in Kansas. D. F. MERRIAM and T. E. KELLY. Map, scale 1″/10 miles.

Montana Bureau of Mines and Geology—Butte, 1960.

Memoir 38. Pre-Beltian Geology of the Cherry Creek and Ruby Mountains Areas, Southwestern Montana. E. W. HEINRICH and J. C. RABBITT. Pp. 40; pls. 5; figs. 9; tbls. 4. Price, $1.00. *While graphite has been mined in the past, only talc is being mined today. Iron may be mined in the near future, and sillimanite deposits are of potential interest.*

Spec. Publication 19. Biennial Report of the Montana Bureau of Mines and Geology, July 1, 1958 to June 30, 1960. Pp. 15; pl. 1. No charge. *In 1959 petroleum production was valued higher than copper production for the third year in a row.*

Bull. 19. Preliminary Report on the Geology and Ground-Water Resources of Northern Blaine County, Montana. E. A. ZIMMERMAN. Pp. 19; pl. 1; figs. 5; tbls. 2. No charge. *Good quality water is available in large amounts from the Flaxville formation.*

New Mexico Geological Survey—Sante Fé, 1960.

Technical Rept. 18. Availability of Ground Water at Proposed Well Sites in Gila National Forest, Sierra and Catron Counties, New Mexico. F. D. TRAUGER. Pp. 20; pls. 3; figs. 2; tbls. 2. *Ground-water resources are sufficient for camp facilities.*

Technical Rept. 19. Ground-Water Levels in New Mexico, 1956. H. O. REEDER and others. Pp. 251; figs. 19; tbls. 52. *In 1956 precipitation was the lowest ever recorded.*

Northern Rhodesia Geological Survey—Lusaka, 1959–1960.

Occasion Paper 8. Some Interpretations Arising from a Remapping of the Katanga System Southeast of Mumbwa, Northern Rhodesia. K. A. PHILLIPS. Pp. 213–223; figs. 2; tbl. 1. *Metallic mineralization is related to contact metamorphism and to regional structures.*

Report 14. Graphite of the Petauke District Eastern Province. A. R. DRYSDALL. Pp. 28; pl. 1; tbls. 11; maps 7. Price, £1 1s. *The presence of several orebodies has been located; however, since the world market for graphite has fallen in recent years, more prospecting is necessary before a final economic assessment is made.*

Ontario Department of Mines—Toronto, 1960–1961.

Sixth Annual Report of the Ontario Fuel Board 1959. Pp. 105; figs. 2; tbls. 13. *Leasing and drilling declined slightly in 1959, but the production of oil and gas increased.*

Vol. LXIX, Pt. 7. Gunflint Iron Range in the Vicinity of Port Arthur. W. W. MOORHOUSE. **Gunflint Iron Formation of the Whitefish Lake Area.** A. M. GOODWIN. Pp. 67; figs. 10; photos 55. 8 Maps, scale 1″/20 miles. *Two papers on the Gunflint Iron Range.*

Vol. LXIX, Pt. 8. Nepheline Syenite Deposits of Southern Ontario. D. F. HEWITT. Pp. 194; figs. 25; photos 39. 2 maps, scale 1/31,680. *Expansion of use of nepheline syenite in glass and ceramic industries results in favorable outlook for future demands.*

Misc. Paper MP-2. Bibliography of Theses in the Precambrian Geology of Ontario. R. M. GINN, Compilator. Pp. 49.

Pennsylvania Geological Survey—Harrisburg, 1960.

Information Circ. 36. The Mineral Industry of Pennsylvania in 1958. R. D. THOMSON, MARY E. OTTE and R. E. ELA. Pp. 47; tbls. 16.

Information Circ. 42. Geologic Interpretation of Certain Aeromagnetic Maps of Bucks, Montgomery, and Lehigh Counties. A. A. SOCOLOW. Pp. 6. *The six quadrangles covered all have simple magnetic contour maps and elicit brief comment.*

Bull. W12. Borehole Geophysical Methods for Analyzing Specific Capacity of Multiaquifer Wells. G. D. BENNETT and E. P. PATTEN, JR. Pp. 25; figs. 8; tbls. 2. *This paper presents the methods of well-logging used to determine the discharge-drawndown relations of individual aquifers supplying a multiaquifer well.*

Bull. M42. Coal, A Petrographic Approach. E. F. KOPPE. Pp. 25; figs. 10; tbl. 1. *Contains an introduction to coal petrography and its potential value to coal utilization of research.*

Somaliland Protectorate Geological Survey—Hargeisa, 1957, 1960.

Rept. 1. The Geology of the Heis-Mait-Waqderia Area, Erigavo District. J. E. MASON and A. J. WARDEN. Pp. 23; maps 5, scale 1/125,000. *Mapping in northeast Somaliland has revealed a two-part basement that is characterized by differences of metamorphism. No evidence of economic mineralization was found.*

Mineral Resources Pamphlet 3. Minerals and Rocks of Hargeisa and Borama Districts. J. L. DANIELS. Pp. 25; figs. 3; tbl. 1. Price, shs. 4/-. *There are promising finds of beryl and columbite.*

Geological Survey of Tanganyika—Dodoma, 1960.

Records of the Geological Survey of Tanganyika, Vol. VIII, 1958. Pp. 113. Price, shs. 17/50. *This volume contains six progress reports, four on regional geology, five on economic geology, and 1 geophysical report.*

Memoir 11. The Rungwe Volcanics at the Northern End of Lake Nyasa. D. A. HARKIN. Pp. 172; pls. 31; tbls. 53; 3 maps, scale 1/250,000. *A systematic study of the geology, history, petrography and petrology of the Rungwe Volcanics that lie in the Western Rift System. This report contributes significantly to the controversy of the chemistry of the volcanics.*

West Virginia Geological Survey—Morgantown, 1960.

Rept. of Investigations 19. Photogeologic Techniques Applied to the Mapping of Rock Joints. VAN NESS D. HOUGH. Pp. 21; pls. 5; figs. 2; maps 5. *A review and expansion of techniques of interpreting geomorphic expressions of jointing and fracturing through the use of aerial photographs.*

A Guide to the Common Fossil Plants of West Virginia. W. H. GILLESPIE and I. S. LATIMER, JR. Pp. 36; pls. 23; figs. 15.

National Science Foundation—Washington, D. C., 1960.

Surveys of Science Resources Series NSF 60-49. Funds for Research and Development in Industry 1957. Pp. 119. Price, 65 cents. *A summary of industrial research and development in relation to financing and company size.*

Surveys of Science Resources Series NSF 60-80. Federal Funds for Science IX. Pp. 89; tbls. 14. Price, 50 cents. *The Federal research and development budget fiscal years 1959, 1960, and 1961.*

U. S. Geological Survey—Washington, D. C., 1960–1961.

Bull. 1071-F. Corals from Well Cores of Madison Group, Williston Basin. W. J. SANDO. Pp. 30; pls. 8; figs. 2. Price, $1.00. *A description of Early Mississippian corals from three wells in northeastern Montana with special reference to their value as facies and time indicators.*

Bull. 1078. Geology and Coal Resources of the Buffalo-Lake De Smet Area, Johnson and Sheridan Counties, Wyoming. W. J. MAPEL. Pp. 148; pls. 23; figs. 6; tbls. 4. *Coal deposits are found in the Wasatch formation; rank ranges between lignite and sub-bituminous.*

Bull. 1081-C. Geology of the Maddux Quadrangle, Bearpaw Mountains, Blaine County, Montana. BRUCE BRYANT, R. G. SCHMIDT, and W. T. PECORA. Pp. 25; fig. 1; map 1, scale 1:31,680. *The quadrangle is in the southeastern part of the Bearpaw Mountains and is underlain by Jurassic and younger sediments and Eocene volcanics. Cause of structural deformation of the volcanics cannot be determined in this quadrangle.*

Bull. 1082-G. Areal Geology of the Little Cone Quadrangle, Colorado. A. L. BUSH, O. T. MARSH, and R. B. TAYLOR. Pp. 423–492; pls. 18–9; figs. 42–46. *Entrada sandstone (Late Jurassic) contains widespread low-grade uranium deposits and vanadium deposits.*

Bull. 1082-F. Geology and Fluorspar Deposits, Northgate District, Colorado. T. A. STEVEN. Pp. 323–422; pls. 12–17; figs. 33–41; tbls. 2. *The largest deposits of fluorspar occur in or adjacent to open spaces along late Tertiary (?) faults.*

Bull. 1083-E. Anomalous Remanent Magnetization of Basalt. ALLAN COX. Pp. 29; figs. 11. Price, 20 cents. *A study of causes of anomalous magnetization with analysis of field and experimental data to show that lightning probably produced anomalous magnetization in basalts in Idaho.*

Bull. 1116-C. Geophysical Abstracts 182, July–September, 1960. J. W. CLARKE, DOROTHY B. VITALIANO, VIRGINIA S. NEUSCHEL, and others. Pp. 167. Price, 40 cents. *Abstract of current literature pertaining to the physics of the solid earth and to geophysical exploration.*

Bull. 1111-B. Field Description and Sampling of Coal Beds. J. M. SCHOPF. Pp. 25–67; pls. 6–27; figs. 3. *Information required in describing coal beds with a discussion of sampling methods.*

Bull. 1111-C,D. Petrology of the Meade Peak Phosphatic Shale Member of the Phosphoria Formation at Coal Canyon, Wyoming, and a Method of X-Ray Analysis for Determining the Ratio of Calcite to Dolomite in Mineral Mixtures. R. A. GULBRANDSEN. Pp. 71–152; pls. 28–33; figs. 4–15; tbls. 19. *The author believes that the mineral mixtures of dolomite, calcite and carbonate-fluorapatite can be explained in part by phase relationships.*

Prof. Paper 282-F. Drainage Basins, Channels, and Flow Characteristics of Selected Streams in Central Pennsylvania. L. M. BRUSH, JR. Pp. 36; pls. 4; figs. 22; tbls. 9. *A study of the influence of the geologic character of drainage basins upon the hydraulic characteristics of 16 natural stream channels at 119 sampling stations.*

Prof. Paper 326. Cenozoic History of Northeastern Montana and Northwestern North Dakota With Emphasis on the Pleistocene. A. D. HOWARD. Pp. 107; pls. 8; figs. 44; tbls. 6; map, scale 1″/5 Miles. *A study emphasizing the Pleistocene history of the north-central Great Plains, with descriptions of glacial and non-glacial deposits and of major drainage changes.*

Prof. Paper 328. Geology of the Yankton Area, South Dakota and Nebraska. H. E. SIMPSON. Pp. 118; pls. 13; figs. 11; tbls. 5; map, scale 1/48,000.

Paleozoic Species in Bairdia and Related Genera. I. G. SOHN. Pp. 105; pls. 6; figs. 15. Price, $1.00.

Prof. Paper 331-A. Archeogastropoda, Mesogastropoda and Stratigraphy of the Ripley Owl Creek, and Prairie Bluff Formations. N. F. SOHL. Pp. 141; pls. 18; figs. 11; tbl. 1. Price, $2.00. *In this report 48 genera and 94 named and 33 unnamed species are discussed and diagnosed; 4 genera are new; 3 new subgenera are proposed; and 21 species and 2 subspecies are described as new.*

Prof. Paper 336. Kyanite, Sillimanite, and Andalusite Deposits of the Southeastern States. G. H. ESPENSHADE and D. B. POTTER. Pp. 120; pls. 11; figs. 58; tbls. 12. Price, $1.25. *A geologic study of deposits of the Al_2SiO_5 minerals in the Piedmont and Blue Ridge provinces.*

Prof. Paper 354-E. Interpretation of the Composition of Lithium Micas. MARGARET D. FOSTER. Pp. 32; figs. 14; tbls. 8. *Relations of lithian muscovites and lepidolites, and of siderophyllites, protolithionites, zinnwaldites, and lepidolites, based on published analyses.*

Prof. Paper 354-F. Zones and Zonal Variations in Welded Ash Flows. R. L. SMITH. Pp. 10; pls. 2. *A concept of zonation in ash flows based on degree of welding and type of crystallization.*

Prof. Paper 356-B. Uranium and Other Trace Elements in Petroleums and Rock Asphalts. K. G. BELL. Pp. 20; fig. 1; tbls. 3. Price, $1.00. *This study presents evidence that petroleum does not act as an ore-forming fluid for uranium, but it may provide a reducing environment in which uranium may be deposited from solutions.*

Water-Supply Paper 1520. Quality of Surface Waters of the United States 1957. Pts. 1–4. North Atlantic Slope Basins to St. Lawrence River Basin. Pp. 641; fig. 1. Price, $2.00.

Water-Supply Paper 1524. Quality of Surface Waters for Irrigation Western United States 1957. S. K. LOVE. Pp. 183; pls. 1. Price, $1.00. *Compilations from 106 network stations.*

Water-Supply Paper 1594-A. Artificial Recharge Through a Well Tapping Basalt Aquifers Walla Walla Area Washington. C. E. PRICE. Pp. 31; figs. 4; tbls. 8. Price, 20 cents. *The water injection caused a rise of water level and increased the amount of ground water in storage, but the subsequent yield and specific capacity of the well were impaired.*

Economic Geology
Vol. 56, 1961, pp. 633–634

SCIENTIFIC NOTES AND NEWS

L. C. NOAKES has been transferred from Canberra, Australia, to London for 2 years as British Commonwealth Geological Liaison Officer, with office in Africa House, Kingsway, London W.C. 2.

A Geochemistry Discussion Group has been formed by those in related fields at Ottawa, with monthly meetings (except in summer). Interested visitors are invited. Information and abstracts of papers are available from Dr. John A. Maxwell, Geological Survey, Ottawa.

ROBIN BRETT has moved from Harvard University to The Geophysical Laboratory, Washington 8, D. C.

THOMAS B. NOLAN has been requested to continue as head of the Geological Survey by the Secretary of the Interior, Stewart L. Udall. Dr. Nolan has been with the Department for 37 years and Director since 1956.

J. M. HARRISON, Director of the Geological Survey of Canada, was elected first president of the newly formed International Union of Geological Sciences at the organizational meeting held in Paris, France, in March. The objectives of the new Union are to promote and encourage the study of geological problems; to facilitate international cooperation in geology and related sciences; and to assist the International Geological Congresses generally held every four years. Thirty-two countries, including United States, Soviet Russia and the United Kingdom have already joined the Union and several more are expected by the end of the year.

Discovery of an isotope of element 103 was announced April 12 by a team of scientists at the Lawrence Radiation Laboratory, operated for the Atomic Energy Commission by the University of California. The Berkeley researchers have suggested the name Lawrencium (chemical symbol Lw), for the new element in honor of the late Ernest O. Lawrence, Nobel prize winner, inventor of the cyclotron and founder of the Laboratory which now bears his name. The discovery was made by nuclear chemists ALBERT GHIORSO, TORBJORN SIKKELAND, ALMON E. LARSH and ROBERT M. LATIMER of the Lawrence Radiation Laboratory. A paper on the research, which was supported by the Atomic Energy Commission, has been sent to the Physical Review for publication.

S. ALEX SCOTT has been appointed chief mining engineer and FRANK C. PICKARD, manager of prospect development in the Mining and Exploration Department of International Minerals & Chemical Corporation. E. C. KRUGER, director, has also announced the appointment of DONALD H. FREAS as geologist and DAVID M. WILLIAMS as mining engineer.

An iron ore discovery in Costa Rica in Guanacaste Province is being evaluated on the 80 square mile concession awarded to International Iron, Inc. of Houston, Texas, and San Jose, Costa Rica. The iron ore content ranges from 45% to 69% Fe and may exceed 400,000,000 tons. Both geophysical and surface geology are being employed in the studies.

CHARLES MILTON, U. S. Geological Survey, has returned from six weeks geological study in Israel. With members of the Israel Geological Survey and the

Hebrew University of Jerusalem he visited the sedimentary copper deposits at Timna, the phosphate mines in the Negev, the oil shale and salt deposits of the Dead Sea region, and the uranium-chromium mineralization in the Chatrurim area.

M. HANS FROHBERG, Consulting Geologist of Toronto, Ontario, was elected President of the Geological Association of Canada for 1961–62 at the annual meeting of the society held in Toronto, March 6, 1961. He succeeds J. M. Harrison, Director, Geological Survey of Canada, Ottawa, Ont.

The Highway Research Board's book on Landslide Investigations (reviewed in this journal, Vol. 53, p. 897) contains a chart by D. J. Varnes on Classification of Landslides. Copies of this excellent chart (18″ × 24″) are now made available at 25 cents each, from Highway Research Board, National Academy of Sciences, 2101 Constitution Ave., Washington, D. C.

GUNNAR KULLERUD of the Geophysical Laboratory in Washington gave a series of ten lectures on "Sulfide Systems" to geology students at Dartmouth College during the winter term.

Dickinson College for the first time has chosen a leader from the earth sciences to receive its Priestley Memorial Award. MAURICE EWING, oceanographer and geologist, became the tenth recipient of the award during Dickinson College's annual Priestley Celebration on March 23.

R. M. DREYER, formerly Assistant Chief Geologist, Reynolds Metals, and Chief Geologist, Kaiser Aluminum, and J. K. BROOKE, Administrative Mining Engineer, Kaiser Aluminum, announce the formation of a consulting practice in mining geology, mining geophysics, and mining engineering with offices at 465 California Street, San Francisco, California.

JOSEPH E. PATCHETT has been awarded the Peacock Memorial Prize (1960) of the Walker Mineralogical Club for his paper entitled, "A Study of the Radioactive Minerals of the Uraniferous Conglomerate, Blind River Area."

TRUMAN H. KUHN, Dean, Graduate Department of Colorado School of Mines and Chairman, Educational Committee A.I.M.E., spoke before the Luncheon Meeting of the Mining and Metallurgical Society of America on "Industry and Mineral Education."

GEORGE E. KRUGER has been promoted to vice president by the Chase Manhattan Bank, New York. Mr. Kruger joined the bank in 1957 as a mining geologist in the petroleum department and was named technical director, mining industry, in 1960. Prior to 1957 he was a consulting mining geologist in South America and in New York.

STEWART H. ROSS is geologist in a team of mining specialists from the Paul Weir Company who are in South Vietnam to assist the Vietnamese government in the development of the Nong Son coal mines, a project sponsored by the U. S. International Cooperation Administration as part of its technical aid program.

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Applications are invited from qualified mineralogists for posts in the Mineralogical Section of the Research and Development Division, Kitwe, which undertakes a wide range of projects in extraction metallurgy for the Rhodesian Companies of the Anglo American Corporation of South Africa Limited. These vacancies are the result of promotions and expansion.

Facilities in the Mineralogical Section include X-Ray diffraction equipment, projection microscope and infrasizing and superpanning equipment as well as the usual microscopes, photographic facilities and sample preparation equipment. Analytical services are available.

Applicants for the post of Mineralogist should have had some experience in determinative mineralogy and in petrology. The Senior Mineralogist, who would report directly to the Assistant Superintendent of the Division, should have sound experience in the above fields and the ability to co-ordinate and supervise the work of a team of three mineralogists. He would also be required to liaise with geologists and research and plant personnel on mineralogical problems. A knowledge of mineral concentration techniques would be an advantage.

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The Secretaries,
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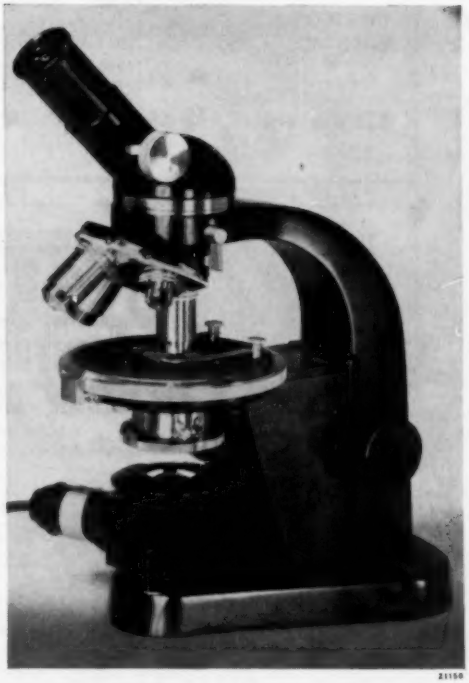

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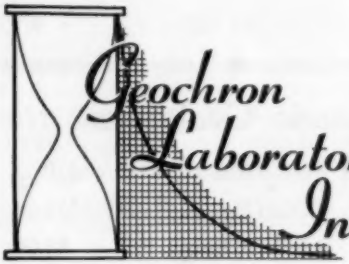

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